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LITHIFICATION OF PLEISTOCENE CLAY AT KAHUKU POINT, OAHU

BY RANDOLPH W. CHAPMAN

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ABSTRACT

At Kahuku Point, on the northern coast of Oahu, Territory of Hawaii, a layer of partially lithified clay rests unconformably on the irregular surface of an elevated marine bench of reef rock and beneath an overlying beach limestone and eolian limestone. The distribution, physical character, and chemical composition of this deposit reveal that it was washed in from areas having a lateritic soil development, deposited subaerially, and subsequently consolidated by calcium and magnesium carbonates leached from the overlying calcareous formations. Field relations indicate that the lithified clay is probably late Pleistocene.

INTRODUCTION

Kahuku Point is a jagged, rocky prominence jutting into the sea and forming the northernmost point on the island of Oahu, Territory of Hawaii (Fig. 1). This remote and desolate peninsula, approximately 900 feet long and 250 feet wide, trends due north. Its surface, which stands about 5 feet above mean sea level, is essentially horizontal, although minutely and intricately dissected by solution into peaks and hollows.

Kahuku Point is carved from an emerged marine bench, composed chiefly of ancient reef rock, which crops out in many places around the margin of Oahu. In

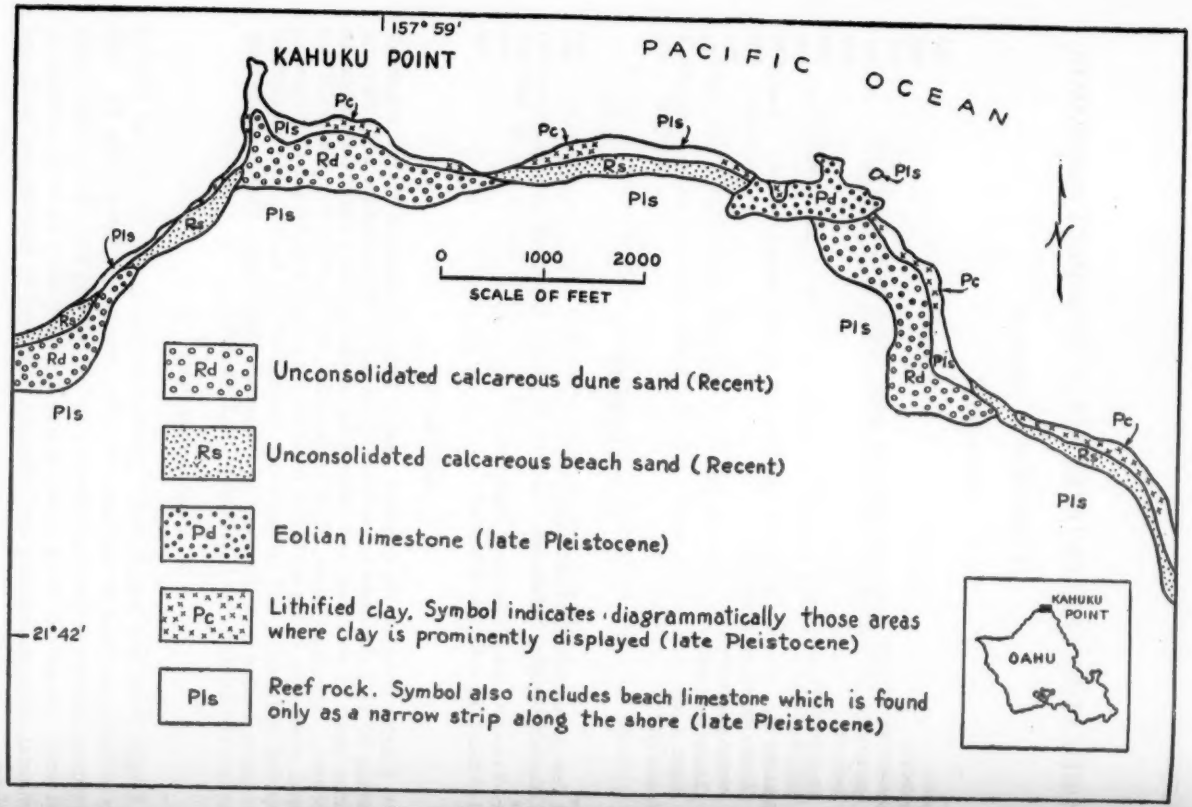

FIGURE 1.—*Geologic map of Kahuku Point, Oahu*
Showing location of lithified Pleistocene clay

the vicinity of Kahuku Point this bench is approximately a mile wide, and it extends as a broad band several miles in either direction along the shore. At the water line, the surface of the bench is approximately 5 feet above mean sea level, but it rises very gently landward. On the south (landward) it is bounded by ancient marine cliffs, cut in the basaltic lavas of the Koolau Mountains, which reveal an old shore line 25 feet above present sea level. On its seaward side, this 5-foot bench is bordered by a lower marine bench, composed of reef rock, which rises to within a few inches of mean sea level and is barely awash at low tide.

The coast line on either side of Kahuku Point is rough and irregular and is characterized by numerous jagged projections protruding into the sea. These projections, composed of emerged reef rock, are separated by broad stretches of white, sandy beach and by long outcroppings of reef rock and beach limestone.

The broad and surging swells of the Pacific Ocean produced by the northeast trade wind beat violently against the northern coast of Oahu (Pl. 1, fig. 1). At Kahuku Point, however, their force is somewhat broken by a living, calcareous reef which extends several hundred feet out to sea and which dissipates each wave into a series of breakers as it moves landward. As a consequence, the air in this region is constantly filled with minute droplets of salt spray.

Weathering and erosion of the Kahuku coast line are chiefly by solution, a process which is rapidly removing the soluble rock. Mechanical abrasion, resulting from calcareous sand being swept back and forth over the rocks by waves, is also a factor. These processes have exposed remnants of a partially lithified clay resting unconformably on the irregular surface of the reef rock and beneath an overlying beach limestone (Pl. 1, fig. 2). The purpose of this paper is to describe the nature of this ancient clay and to explain its origin.

FIGURE 1.—*Geologic map of Kahuku Point, Oahu*
Showing location of lithified Pleistocene clay

The writer is indebted to Professor Harold S. Palmer for his valued advice in the field and for his critical reading of the manuscript. He is also greatly obliged to Dr. G. Donald Sherman for making and interpreting the chemical analyses of the clay and related rocks. Dr. Gordon A. Macdonald assisted the writer in the preparation of thin sections, and Dr. Chester K. Wentworth generously permitted the use of his laboratory for this work. Mr. Bernard F. Chapman drafted Figures 1 and 2. Photographs were taken by the Photographic Section of the 30th Bombardment Group (Heavy) commanded by Colonel E. Patteson.

GENERAL GEOLOGIC RELATIONS

STRATIGRAPHY

Figure 1, modified after the geologic map prepared by Stearns (1939, Pl. 2), shows the disposition of rocks and sediments along the coast near Kahuku Point. The writer has modified slightly some of Stearns' boundaries and has added crosses to indicate localities where the lithified clay is most prominently displayed. Since the outcrops of this clay are scattered and discontinuous, the crosses merely indicate its general distribution.

A brief description of each formation, together with its age, thickness, and stratigraphic relationship, is given in Table 1. Both the reef rock and the beach limestone

belong to that group which Stearns and Vaksvik (1935, p. 165-169) have termed "consolidated calcareous marine sediments." The eolian limestone is equivalent to their "consolidated calcareous dunes" (Stearns and Vaksvik, 1935, p. 169). The thicknesses shown in Table 1 were measured on the formations in the vicinity of Kahuku Point only and do not necessarily represent the maximum thicknesses as displayed elsewhere on the island.

The reef rock is obviously the oldest since it forms the basement on which the other rocks rest. It has been dated late Pleistocene by Stearns and Vaksvik (1935, p. 166) and by Stearns (1945, p. 1074), but this dating has not been accepted by all geologists. Stearns and Vaksvik (1935, p. 169) have also determined that the eolian

TABLE 1.—*Stratigraphy at Kahuku Point, Oahu*

Name	Nature	Age	Thickness in Feet
Unconsolidated calcareous dune sand	Fine-grained, cross-bedded, cream-colored sands. Conglomeratic in places.	Recent	0 to 15±
Unconsolidated calcareous beach sand	Fine-grained, cream-colored sands, consisting of coral, shells, algae, *etc.*	Recent	0 to 20
Eolian limestone	Thin-bedded and cross-bedded consolidated calcareous dunes.	Late Pleistocene (Waipio stand)	0 to 15±
Beach limestone	Cream-colored lithified calcareous beach sand like present beach sands.	Late Pleistocene (Waipio stand)	0 to 7
Lithified clay	Yellowish-brown and reddish-brown, fine-grained clay in various stages of lithification.	Late Pleistocene (Waipio stand)	0 to 1
Reef rock	Coral heads and coralline algae with some clastic material.	Late Pleistocene	135

limestone (consolidated calcareous dunes) is probably late Pleistocene, and state that it is older than the Waimanalo stand of the sea (plus 25 feet) and mostly correlative with the Waipio stand (minus 60 feet). It will be shown subsequently that the beach limestone is contemporaneous with the eolian limestone. Since the lithified clay lies beneath the beach limestone, and, in places, beneath the eolian limestone, it is older than these formations. Thus the writer believes that it was deposited during the Waipio stand of the sea (late Pleistocene) and shortly before the overlying calcareous rocks. However, more positive evidence of the age of Oahu's coral reefs is necessary before the age of the clay can be verified. The unconsolidated calcareous beach sand and the unconsolidated calcareous dune sand are unquestionably Recent.

Inasmuch as the reef rock, the beach limestone, and the eolian limestone are all

FIGURE 1. VIEW EAST OF KAHUKU POINT, OAHU
Looking east; showing effects of wave action on coast line. In right foreground is lower wave-cut bench. In background is eolian limestone resting on reef rock. Note stack of eolian limestone in right background.

FIGURE 2. VIEW SOUTHWEST OF KAHUKU POINT, OAHU.
Looking northeast; showing lithified clay (dark layer) lying unconformably on reef rock and beneath beach limestone. Hammer handle points to upper contact of clay. Note crusty type of clay near water line and loose, earthy type farther inland. In left background is Kahuku Point, carved from 5-foot bench, with overlying calcareous dunes.

WAVE ACTION AND LITHIFIED CLAY AT KAHUKU POINT, OAHU

FIGURE 1. VIEW SIMILAR TO THAT IN PLATE 1, FIGURE 2.
Showing crusty type of lithified clay (dark material) lying unconformably on reef rock and beneath beach limestone. Hammer handle points to upper contact of clay.

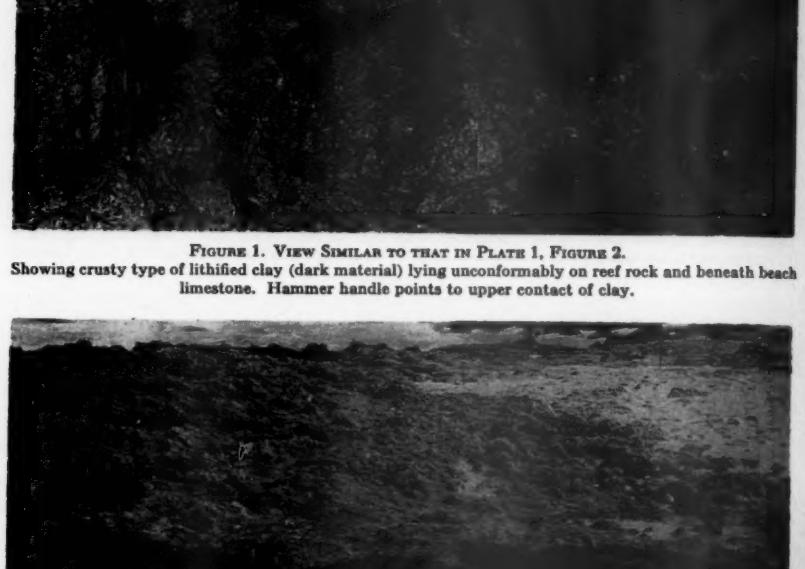

FIGURE 2. VIEW SIMILAR TO THAT IN PLATE 1, FIGURE 2.
Showing loose earthy type of lithified clay (dark gray) resting unconformably on reef rock (light gray). Hammer handle points to base of overlying beach limestone which is pock-marked by solution. Much of the loose clay has been removed by wave action.

LITHIFIED CLAY AT KAHUKU POINT, OAHU

pertinent to the present problem, they will be described in some detail before the lithified clay itself is discussed.

REEF ROCK

The clastic sediments in the vicinity of Kahuku Point rest unconformably on the surface of an emerged fringing reef (Pl. 1, fig. 2) which, at present, stands 5 feet above mean sea level and which, according to Stearns and Vaksvik (1938, p. 195), is 135 feet thick. It crops out as an interrupted band of varying width around Oahu. The existing surface is probably not that of the original reef but represents an erosion plane which was cut when sea level was higher.

The unconformity between the reef rock and the overlying formations is irregular. In general it is approximately 5 feet above mean sea level, *i.e.* it coincides essentially with the present topographic surface of the reef. Locally, however, it either dips below sea level or rises so high that it intersects the present topographic surface. The unconformity is also irregular in detail because the overlying sediments were deposited on a rough, jagged surface (Pl. 1, fig. 2; Pl. 2).

Irregularities in the reef surface have been further exaggerated along the shore by wave action. Pinnacles, pits, stacks, and caverns indicate rapid solution by the ever-surging sea water. Near Kahuku Point, solution pits up to 6 feet across and 6 feet in depth have been observed.

Lithologically, the emerged reef consists chiefly of coral heads and coralline algae with a variety of clastic, calcareous material filling the interspaces. In places, near its upper surface, the reef has been recrystallized to a dense, cream-colored mosaic by calcium carbonate which has leached downward from the overlying eolian limestone.

BEACH LIMESTONE

Lying unconformably on the irregular surface of the reef rock is a poorly stratified, well-lithified beach limestone (Fig. 1; Pl. 1, fig. 2; and Pl. 2). This rock forms a narrow, nearly continuous band, fringing the shore east and southwest of Kahuku Point. Due to its narrow width and to the fact that it is covered in many places by unconsolidated calcareous beach sand, this rock could not be delineated on the geologic map, and, accordingly, on Figure 1 it is not differentiated from the reef rock. Stratification in the beach limestone is not particularly obvious in most places, but where observed it is essentially horizontal. In general, the thickness ranges up to 3 feet. However, approximately half a mile southwest of Kahuku Point, an exceptionally fine exposure attains a thickness of 7 feet, and the stratification dips about 5° seaward.

This formation clearly represents an ancient beach sand formed, presumably, while the sea was retreating from one of its higher stands. Like the reef rock, the beach limestone is being destroyed rapidly by solution, and its surface is pock-marked by numerous peaks and hollows (Pl. 2, fig. 2).

Petrographically, this deposit consists "of cream-colored to white cemented water-worn grains of comminuted marine organisms, in most places containing a few shells and fragments of water-worn coral" (Stearns and Vaksvik, 1935, p. 168). The

grains are somewhat coarser than those in the eolian limestone and range in diameter from 0.5 mm. to 1.5 mm. At its base, the beach limestone is conglomeratic with fragments of coral and shell 1 to 2 inches in diameter and, in one place, boulders 6 to 10 inches across. At its base, also, the rock is reddish brown in places due to iron-oxide particles.

The lithification of this formation will be discussed in a later section.

EOLIAN LIMESTONE

A mile due east of Kahuku Point an elongate outcrop of eolian limestone 1800 feet long and 400 feet wide (Fig. 1; Pl. 1, fig. 1) is well laminated and ranges from 12 to 15 feet in thickness. Cross bedding is clearly evident and dips 35°N. Solution has honeycombed this formation causing a dense, reticulate pattern of peaks and hollows. Close inspection of fresh samples of the eolian limestone shows that it is porous and loosely cemented in its upper portion, but somewhat more firmly cemented near its base. Its component grains of coral and shell material are generally smaller and more uniform in size than those of the beach limestone and are much better rounded. They range from 0.5 to 0.75 mm. in diameter.

In some places the eolian limestone rests unconformably on the reef limestone and on masses of lithified clay, whereas at other points it apparently grades downward into the poorly stratified beach limestone already described. The change from a well-laminated, cross-bedded, porous rock to a coarser-grained formation without conspicuous structure is gradual and complete. It indicates that the beach limestone and the eolian limestone are essentially contemporaneous, although differing structurally. This paradox can be easily and logically explained, however, if we assume that a considerable area of calcareous beach sand was formed in this region by an ancient, receding sea and that subsequently the wind reworked its upper portion into calcareous dunes.

Eolian limestone is not found elsewhere within the region under consideration except in a locality half a mile southwest of Kahuku Point, where a 6-inch layer of eolian limestone rests on a thick exposure of beach limestone, separated by a gradational boundary. This outcrop is too small to be shown in Figure 1.

The lithification of this formation will be discussed in a later section.

LITHIFIED CLAY

DISTRIBUTION AND GENERAL CHARACTER

The lithified clay is a yellowish-brown to reddish-brown, fine-grained material interposed between the eroded reef rock and the overlying beach limestone and eolian limestone (Pl. 1, fig. 2; Pl. 2). It is exposed only along the shoreline east and southwest of Kahuku Point, for only along here is the contact between the reef rock and the overlying beach limestone and eolian limestone exposed. An exceptionally good exposure was found a quarter of a mile southwest of Kahuku Point near the "Kahuku Kabana." The outcrops of the clay are somewhat sporadic; at some points they are abundant, whereas at others there are none. Nevertheless, the distinctive character and restricted vertical distribution of the rock make it a

convenient horizon marker. Figure 1 indicates some of the principal localities where the clay may be studied to best advantage. However, the writer has found outcrops of lithified clay half a mile beyond the eastern edge of this area.

The clay has two contrasting lithologic modifications: (1) soft, poorly lithified, earthy material, and (2) hard, crusty masses. Gradations between these two extremes occur. The earthy material constitutes a distinct layer, without stratification, between the underlying reef rock and the overlying beach limestone (Pl. 1, fig. 2; Pl. 2, fig. 2). The clay varies considerably in thickness because it was deposited on the irregularly eroded surface of the reef rock. In a depression in the reef it may be 12 inches thick, whereas on the tops of hummocks it may be very thin or absent. Several excellent exposures of the contact, where the surface of the reef is even and horizontal, showed the thickness of the clay to be about 6 inches. This loose, earthy material is found only at a distance from the water's edge out of reach of the most intense wave action.

The hard, crusty type of clay exists as irregular, spiny masses in the reef rock, generally near the water's edge where the waves are dissolving the reef (Pl. 1, fig. 2; Pl. 2, fig. 1). The crusty masses are extremely irregular in shape, and range from 2 inches to 2 feet across. Some masses, slightly more resistant to solution than the reef rock, are perched on top of pinnacles of the reef rock, where they resemble irregular concretions or perched boulders. Other masses have assumed the shape of a bowl or saucer; their configuration obviously was determined by a solution pit in the reef rock. Some of the crusty masses lie as much as 18 inches below the general level of the reef surface.

In many places the upper surface of both clay types grades into the overlying beach limestone within a distance of approximately 1 inch. The lower boundary, however, adjacent to the reef, is everywhere sharp and distinct and in some exposures is marked by a hard, black shell, a millimeter or less in thickness.

PETROGRAPHIC NATURE

Since both the soft, earthy type and the crusty type are varieties of the same clay, the petrography of both can be discussed simultaneously.

The poorly lithified, earthy clay is yellow to yellowish brown. The crusty type is also yellowish brown on the freshly broken surface, but weathered surfaces are dark, reddish brown. These colors suggest subaerial accumulation. Both clay types are even-textured and very fine-grained and show various degrees of compactness. The writer has observed neither sand grains nor pebbles in that clay shown within the area of Figure 1, but half a mile beyond the eastern boundary he found small limestone pebbles 2 mm. in diameter and fragments of coral 2 to 3 inches across.

In thin section the clay is chiefly a confused mass of very fine grains of limonite, carbonates, and opaque, unidentifiable matter. The relative percentages of these constituents vary considerably in different specimens. For example, the crusty type of clay contains more carbonates and is firmly cemented by them. In contrast, the soft, poorly lithified type contains less carbonates and is a loosely cemented, impalpable powder, the greatest lithification being in a zone adjacent to the overlying limestone. The sizes of the grains in both types are the same.

Under the microscope the crusty type consists mostly of limonite grains and of particles of unidentifiable matter, grouped together with carbonates into irregular aggregates. Interspersed between these aggregates are some open pore spaces and

TABLE 2.—*Chemical analyses of lithified clay and related rocks from Kahuku Point, Oahu*

	Heated to 110°C	Strong HCl Extraction		Soluble in Strong HCl			
Sample No.	H_2O and other volatiles Per cent	Insoluble fraction Per cent	Soluble fraction Per cent	R_2O_3 Per cent	$CaCO_3$ Per cent	$MgCO_3$ Per cent	Total carbonates Per cent
1	3.7	18.1	78.2	5.8	66.8	5.6	72.4
2	2.4	19.7	77.9	7.4	66.3	4.2	70.5
3	2.8	25.4	71.8	5.6	61.8	4.4	66.2
4	4.3	34.3	61.4	9.4	48.3	3.7	52.0
5	6.6	32.0	61.4	9.0	51.0	1.4	52.4
6	16.8	54.5	28.7	20.4	7.9	0.4	8.3
7	1.2	1.2	97.6	0.4	96.3	0.9	97.2
8	0.5	0.5	99.0	0.6	97.1	1.3	98.4
9	0.5	2.4	97.1	2.0	94.8	0.3	95.1
10	0.9	1.6	97.5	0.8	96.0	0.7	96.7
11	0.2	1.1	98.7	0.8	97.6	0.3	97.9
12	24.9	52.6	22.5	21.8	0.5	0.2	0.7

1. Typical crusty type of lithified clay.
2. Typical crusty type of lithified clay. Sample taken from clay close to its contact with underlying reef rock.
3. Lithified clay somewhat less consolidated than typical crusty type. Sample taken from clay close to its contact with underlying reef rock.
4. Poorly lithified clay from near gradational contact with overlying beach limestone.
5. Poorly lithified clay not immediately adjacent to either upper or lower contact.
6. Soft, very poorly lithified clay from interior portion of clay layer.
7. Reef rock from a point 3 inches below contact with overlying lithified clay.
8. Reef rock from a point immediately adjacent to contact with overlying lithified clay.
9. Similar to sample 8.
10. Beach limestone from a point immediately adjacent to gradational contact with underlying lithified clay.
11. Beach limestone from a point about 4 inches above contact with underlying lithified clay.
12. Lateritic soil from Poamoho region about 2 miles north of Wahiawa, Oahu.

Notes:

1. All specimens, except No. 12, were collected from a point a quarter of a mile southwest of Kahuku Point where the relationships of the lithified clay, the beach limestone, and the reef rock are particularly clear.
2. All analyses were made by Dr. G. Donald Sherman, Hawaiian Agricultural Experiment Station, University of Hawaii, Honolulu, T. H.

also numerous areas and stringers of fine-grained, recrystallized carbonates, apparently introduced secondarily into open spaces. The introduced carbonates which convert the clay to its hard crusty form are less abundant in the loose earthy type.

CHEMICAL COMPOSITION

The fine grain of the lithified clay precludes the determination by microscopic means of its mineral composition, and any consequent inference as to its chemical composition. Accordingly, it was necessary to resort to chemical analyses of the clay. Table 2 contains several analyses of lithified clay, reef rock, and beach limestone made by Dr. G. Donald Sherman of the Hawaiian Agricultural Experiment Station. In Figure 2 the clay analyses are plotted on a variation diagram. The table and diagram reveal a number of significant facts:

(1) The soft, poorly lithified clay contains more water than the hard, crusty type, probably because of its greater porosity. Other volatiles are present only in small amount.

(2) The ratio of soluble matter (chiefly calcium carbonate and magnesium car-

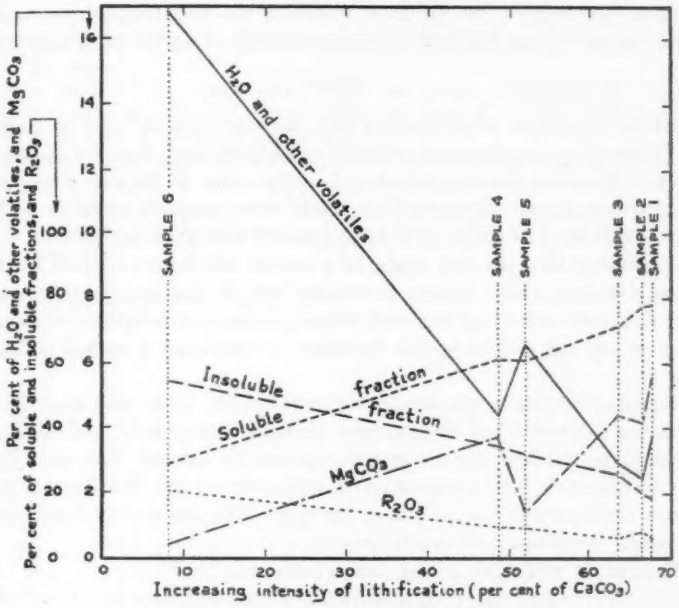

FIGURE 2.—*Variation diagram of lithified clay*
Showing relationship of chemical composition to intensity of lithification

bonate) to insoluble matter is roughly directly proportional to the degree of litbification. The insoluble matter consists chiefly of clay-sized particles of silicate minerals, secondary clay minerals, and heavy minerals. Traces of organic matter are also present in the insoluble portion. Whether the magnesium carbonate in the soluble fraction is magnesite or dolomite is not known.

(3) The soluble oxides (R_2O_3) show an irregular proportional decrease as the clay becomes more firmly lithified. They comprise approximately 80 per cent of Fe_2O_3, 19 per cent of Al_2O_3, and 1 per cent of P_2O_5, according to further work by Sherman.

(4) The reef rock and the beach limestone consist chiefly of calcium carbonate with small quantities of insolubles, soluble oxides (R_2O_3), and magnesium carbonate.

(5) Of all the samples that have been studied, both in the laboratory and in the field, No. 6 is most like a lateritic soil. The chemical similarity may be readily observed by comparing the analysis of No. 6 with that of No. 12 which is a lateritic soil from the Poamoho region about 2 miles north of Wahiawa, Oahu. No. 12 contains more water and other volatiles than No. 6 but is correspondingly lower in

calcium carbonate. In No. 6 the calcium carbonate has displaced a portion of the water in the pore spaces.

Thus, the chemical evidence indicates that the clay has been derived from a lateritic soil and has been lithified through the introduction of calcium and magnesium carbonates into the pore spaces.

Chemical tests suggest that the dark, thin layer that in some places separates the clay from the underlying reef rock is composed chiefly of a series of manganic oxides.

ORIGIN

To explain the origin of the lithified clay, three hypotheses must be considered.

The lithified clay may have accumulated as a marine deposit on the surface of the reef rock. However, the red, brown, and yellow colors of the clay strongly imply subaerial accumulation. Moreover, the thin layer of manganic oxides at the base of the clay is difficult to explain by a hypothesis of submarine deposition.

The possibility that the clay might be a residue which accumulated as the reef rock was dissolved away appears untenable because the boundary between the lithified clay and underlying reef rock is always sharp and distinct. There is no evidence of any solution along this boundary; it resembles a normal sedimentary contact.

In the writer's opinion, the lithified clay at Kahuku Point was washed in from areas having a lateritic soil development, deposited subaerially, and subsequently consolidated by calcium and magnesium carbonates leached from overlying calcareous formations. This explanation is supported by the distribution, physical character, and chemical composition of the clay. The sequence of events involved in the origin is summarized briefly below:

(1) Growth of reef rock during late Pleistocene.

(2) Elevation of reef above sea level in late Pleistocene, and erosion and solution of its surface by retreating sea and by rainwash.

(3) Lateritic clay, derived from weathered basalt of the Koolau Range to the south, was spread subaerially upon this surface by streams. The slightly acid waters of the streams carried manganese in solution, and as they leached through the freshly deposited lateritic clay to the reef rock below they became neutral and precipitated their manganese as a thin layer of manganic oxides.

(4) Deposition of a layer of calcareous beach sand over the outer fringe of the clay apron, thus protecting the latter from erosion.

(5) Retreat of sea to its Waipio stand (minus 60 feet), and shifting of the upper portion of the calcareous beach sands by the wind to form dunes.

(6) Beginning of encroachment of sea toward its Waimanalo stand (plus 25 feet), and simultaneous consolidation of calcareous beach sand, calcareous dune sand, and underlying clay by the following processes: (a) Carbonated rain water, sinking through the beach sand and dune sand, dissolved calcium and magnesium carbonates from the upper portions and redeposited them below, thus cementing the lower portions of both formations as well as portions of the underlying clay. (b) Dashing of breakers against the partly consolidated rocks released carbon dioxide from the resulting spray droplets and caused precipitation of calcium and magnesium

carbonates in pore spaces. The importance of the latter process is indicated by the fact that the well-lithified, crusty type of clay is found only near the water's edge. The lithification of beach sediments by processes similar to these has been described by Branner (1904, p. 171-196) and by Daly (1924, p. 135-140).

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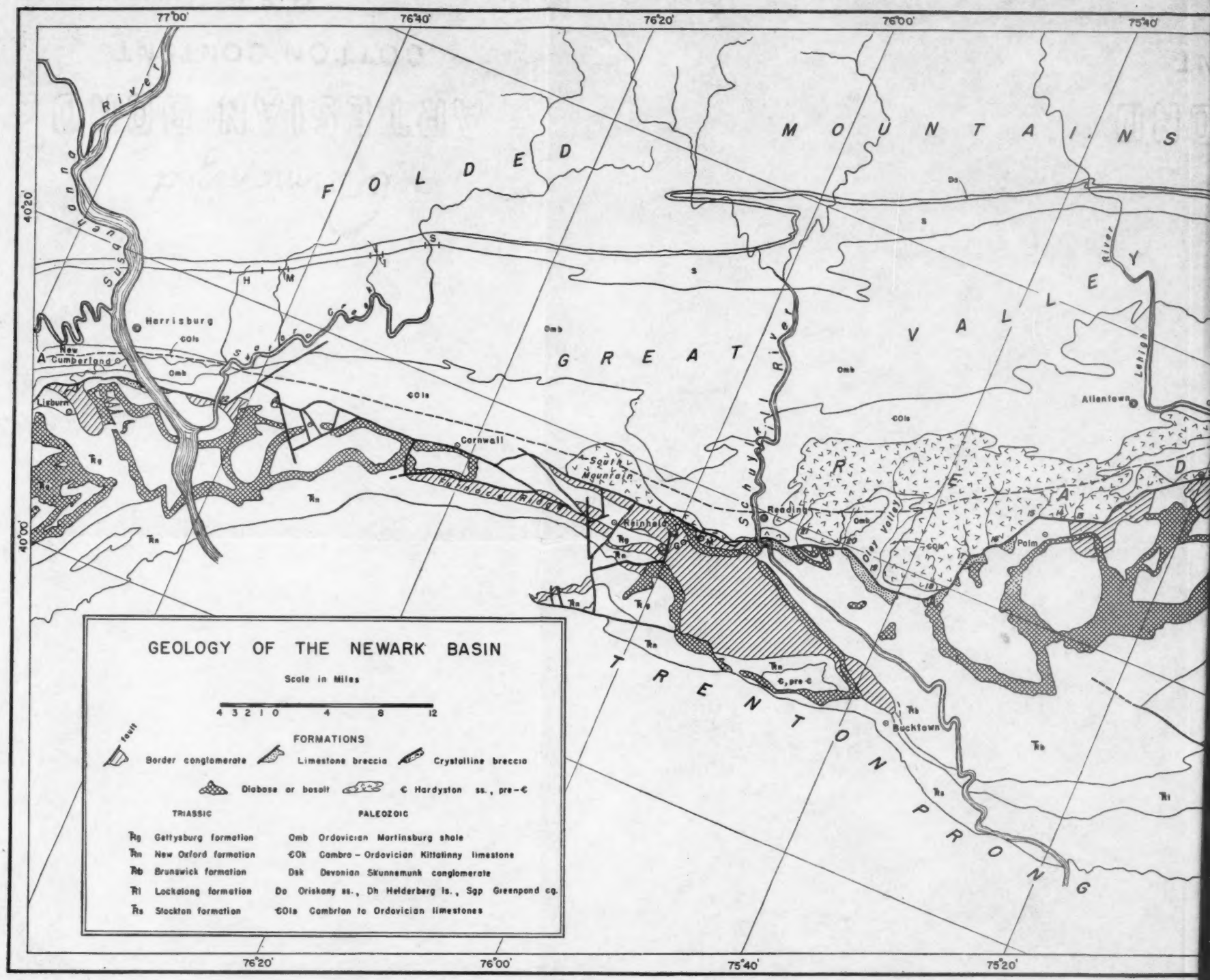
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76°40'
76°20'
76°00'
75°40'
40°20'
40°00'
Susquehanna River
FOLDED MOUNTAINS
GREAT VALLEY
Schuylkill River
Lehigh River
Harrisburg
New Cumberland
Lisburn
Cornwall
South Mountain
Reading
Allentown
Palm
Blue Mountain
TRENTON
PRONG
Buchtown
H
M
S
Dp
Dh
Omb
COls
Trg
Trn
Trb
Trl
Trs
C, pre-C
GEOLOGY OF THE NEWARK BASIN
Scale in Miles
4 3 2 1 0 4 8 12
fault
FORMATIONS
Border conglomerate
Limestone breccia
Crystalline breccia
Diabase or basalt
C Hardyston ss., pre-C
TRIASSIC
Trg Gettysburg formation
Trn New Oxford formation
Trb Brunswick formation
Trl Lockatong formation
Trs Stockton formation
PALEOZOIC
Omb Ordovician Martinsburg shale
COk Cambro – Ordovician Kittatinny limestone
Dsk Devonian Skunnemunk conglomerate
Do Oriskany ss., Dh Helderberg ls., Sgp Greenpond cg.
COls Cambrian to Ordovician limestones
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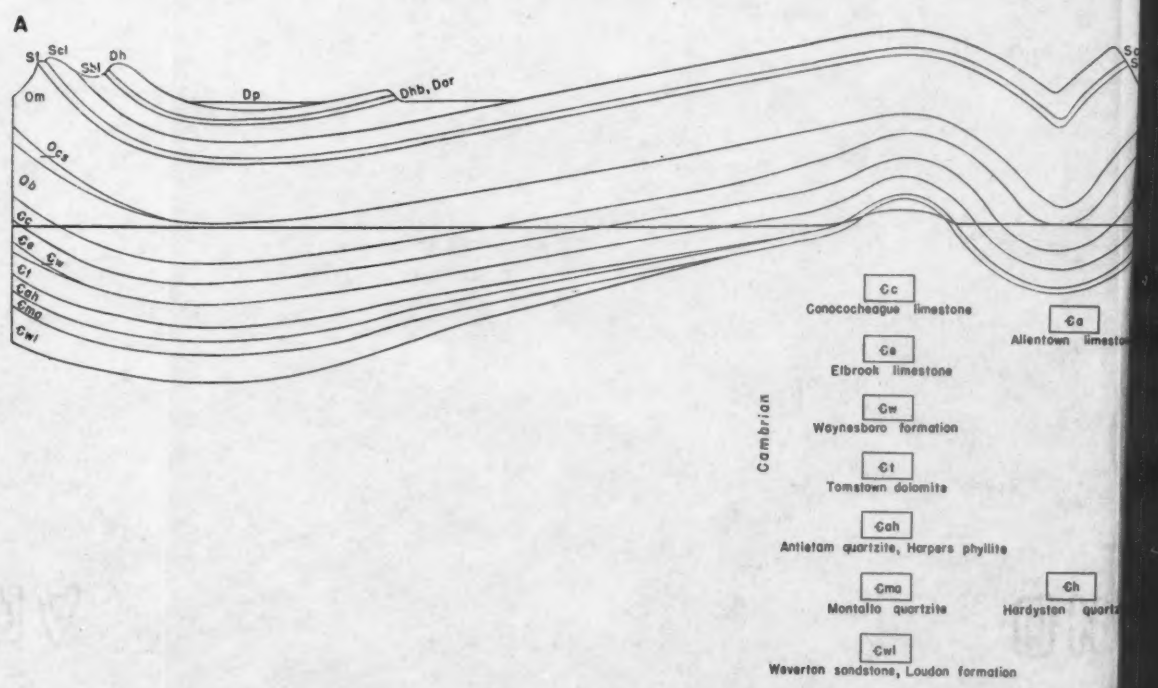
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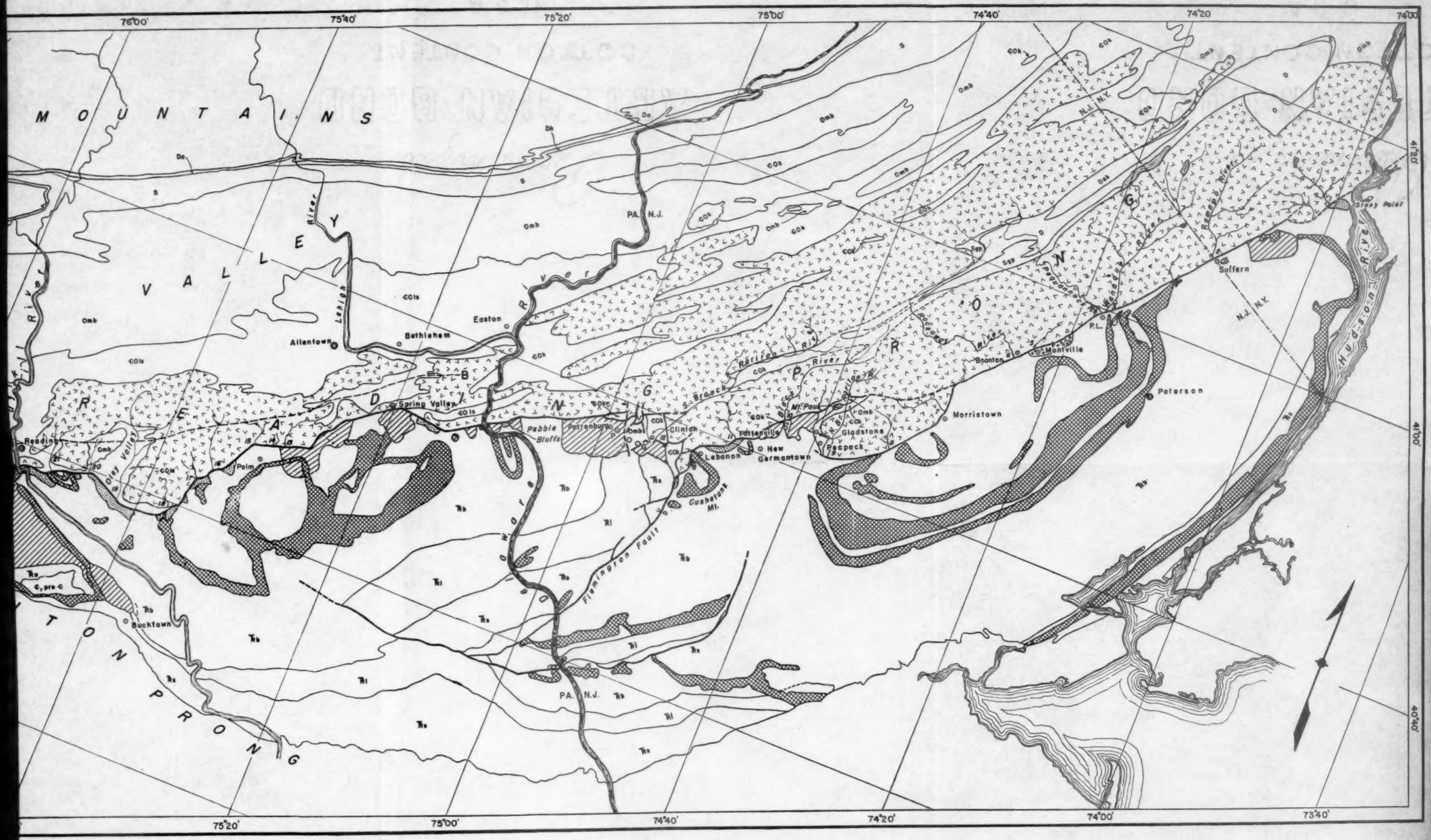
Cambrian
Cc Conococheague limestone
Ce Elbrook limestone
Cw Waynesboro formation
Ct Tomstown dolomite
Cah Antietam quartzite, Harpers phyllite
Cma Montalto quartzite
Cwl Waverton sandstone, Loudon formation

GEOLOGY OF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MOUNTAINS
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PRONG
Delaware River
Lehigh River
Schuylkill River
Hudson River
Raritan River
Passaic River
Ramapo River
Pequannock River
Wanaque River
Pompton River
Second River
Saddle River
Ramapo Fault
Flemington Fault
Cushetunk Mt.
Sourland Mts.
Stony Point
Suffern
Monsey
Spring Valley
Nanuet
Haverstraw
Peekskill
Tomkins Cove
Buchanan
Verplanck
Peekskill
Oscawana
Croton Point
Ossining
Tarrytown
Irvington
Dobbs Ferry
Hastings
Yonkers
Mount Vernon
New Rochelle
Pelham
Mamaroneck
Port Chester
Rye
Harrison
White Plains
Scarsdale
Bronxville
Eastchester
Tuckahoe
Mount Kisco
Bedford
Pound Ridge
Greenwich
Stamford
Darien
Norwalk
Westport
Wilton
Ridgefield
Danbury
New Canaan
Redding
Easton
Bethlehem
Allentown
Reading
Palm
Oley Valley
Pebble Bluffs
Pottersville
Clinton
Lebanon
New Germantown
Mt. Paul
Gladstone
Pecpack
Morristown
Boonton
Montville
Peterson
Paterson
Buckhorn
Buckhill
Buckhill Falls
Buckhill Falls
Buckhill Falls

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APPALACHIAN DRAINAGE AND THE HIGHLAND BORDER SEDIMENTS OF THE NEWARK SERIES

BY CHARLES W. CARLSTON

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ABSTRACT

The highland border fanglomerates of the Newark basin in New York, New Jersey, and eastern Pennsylvania show no extraordinary correlation with present drainage either in distribution or lithologic character and degree of rounding of their gravels. The writer found no evidence of deposition of any of the fanglomerates by major streams and no evidence that any of the present streams enter the basin through Triassic-filled remnants of Triassic valleys.

Available evidence indicates that streams which deposited the Newark fanglomerates were relatively short and steep, consequent on the northwest border fault scarp or flexure. Variations in lithologic character of the fanglomerates were due largely to the rock types exposed along the margin of the northwest highland block. The lithologic character of the Newark sediments and particularly of the basal Stockton supports this hypothesis.

INTRODUCTION

PURPOSE OF PRESENT INVESTIGATION

Meyerhoff and Olmsted (1936) consider that major drainage lines across the folded Appalachians were essentially inherited from Permian time because, as a result of Appalachian orogeny, the drainage divide of the central and northern Appalachians was located northwest of Kittatinny Ridge. Streams heading along this watershed established courses southeastward to the sea by utilization of structural sags, by superposition from asymmetric folds and low-angle overthrust sheets, and by subsequent erosion along fault zones transecting resistant rock barriers. The major drainage lines were established before Newark time and have remained essentially unmodified in the Blue Ridge, Reading Prong, and Piedmont areas.

The hypothesis was developed partly on the basis of the following lines of evidence: (1) close correlation between present drainage lines and Triassic conglomerates; (2) Triassic conglomerates associated with major rivers contain well-rounded pebbles of middle and upper Paleozoic formations now found only many miles back in the folded Appalachians; (3) Triassic conglomerates associated with present minor rivers contain pebbles of rock types now exposed within a short distance of the Triassic basin; (4) present streams flowing southeastward across the Triassic basin enter the basin through V-shaped re-entrants in the highland border which contain Triassic sediments.

A significant aspect of the Meyerhoff-Olmsted hypothesis is the use of Newark sediments as a clue to the drainage conditions in the northern and central Appalachians in late Triassic time. If a study of Newark sediments can lead to any valid

deductions concerning the paleogeography of the terrain north and northwest of the Triassic basin in Newark time and the types of streams draining that terrain, a new method is provided for judging the validity of hypotheses on the origin of Appalachian drainage.

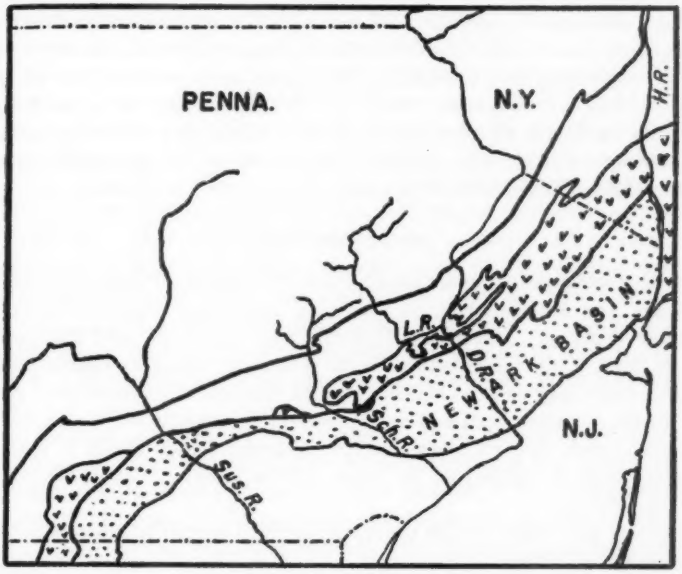

FIGURE 1.—*Index map*

The present paper contains field observations and conclusions from the writer's studies of the western border sediments of the Newark series.

METHODS OF WORK

The writer visited every area of Triassic border fanglomerate between the Hudson River and the northeast end of the Carlisle Prong, or the Blue Ridge (Fig. 1). Three general lines of investigation were followed in the field work.

(1) A study was made of the degree of matching of Triassic fanglomerates with stream valleys in the highland areas. This included mapping of the fanglomerates, the results of which comprise Plate 1.

(2) The physical character of the pebbles, cobbles, and boulders in the fanglomerates was studied. At each outcrop the dimensions of the largest cobble or boulder seen were measured and recorded. Degree of rounding was recorded in empirical terms: rounded, subrounded, curvilinear, subangular, and angular. One or more pebble counts at each outcrop were made by recording the number of pebbles or cobbles of different lithologic character in 1 square yard of outcrop. Variations in character of the deposit, such as amount and thickness of interbedded sandstone or shale, were recorded. Thickness of the section was measured or estimated, and

dip and strike were recorded. Finally, a careful search was made for pebbles or cobbles containing fossils or of distinctive lithology which could be definitely correlated with a stratigraphic source.

(3) Exhumed Triassic valleys or re-entrants containing Triassic sediments were sought. Wherever a river enters the Triassic basin the writer looked for outcrops of Triassic sediments for at least half a mile back of the western border fault. A reconnaissance search was made farther back in each valley. In this the writer was aided by Meyerhoff, who provided a list of nine localities where the presence of re-entrants had been inferred. When no evidence proving the existence of such re-entrants was found, Doctors Meyerhoff and Olmsted and the writer had a 3-day field conference, which, while essentially negative in results, gave assurance that no important field relations were overlooked.

ACKNOWLEDGMENTS

The paper was prepared under the direction of Professors Douglas Johnson, Walter H. Bucher, and Marshall Kay, of Columbia University. Howard A. Meyerhoff, Lawerence Whitcomb, and Arthur N. Strahler have critically read the manuscript. Much of the text was presented before the Graduate Seminar in Geomorphology at Columbia University and has benefited by the criticism of its members. Responsibility for the interpretations and conclusions rests, however, solely with the writer. During field work the writer was aided by conferences with Henry B. Kümmel, B. L. Miller, Ralph L. Miller, R. W. Stone, Girard Wheeler, George W. Stose, and Meredith Johnson.

For typing and editing of this paper, the writer is especially indebted to his wife, Catherine Ball Carlston.

NATURE OF THE TRIASSIC SEDIMENTS

GENERAL ASPECT

In New York, New Jersey, and eastern Pennsylvania the Newark series is divided into three units—the Stockton, Lockatong, and Brunswick formations. The Stockton, basal member, consists of 2300 to 6700 feet of coarse arkose conglomerate and yellow micaceous, feldspathic sandstone overlain by brown-red sandstone and soft red argillaceous shale. The Lockatong formation consists of 1700 to 3600 feet of hard, dark shale; dark-gray and green sandstone with interbedded carbonaceous shale; hard, massive argillite, dark-red shale; and some thin layers of calcareous shale. The Lockatong formation tongues out to the west at the Schuylkill and has not been recognized beyond that point. The Brunswick gradationally overlies the Lockatong and consists of up to 12,000 feet of red shale with interbedded sandstone layers and some interbedded purple, green, yellow, and black shale. Excellent descriptions of the Newark series have been published by Kümmel (1897; 1898; 1899; 1900) and Lewis and Kümmel (1940).

The New Oxford formation is the equivalent of the Stockton formation west of the Schuylkill. In the Fairfield-Gettysburg folio Stose and Bascom (1929, p. 9) have recognized a lower 3000 feet, in which quartz conglomerates and arkose beds are more

common, and an upper 4000 feet of soft red shale and sandstone. Near the middle of the formation 500 feet of the section is characterized by thicker and more numerous white to gray-white to gray micaceous sandstone beds. The Gettysburg formation is the equivalent of the Brunswick formation west of the Schuylkill. In the Gettysburg-Fairfield area it consists of about 16,000 feet of red shale and soft red sandstone and minor amounts of dark impure limestone, black carbonaceous to green and yellow shale, and white sandstone (Stose and Bascom, 1929, p. 9, 10).

Along the northwest border of the Triassic basin occur closely spaced deposits of conglomerate or breccia which range from less than half a mile in greatest diameter to many miles in width and length. Nearly all are in the Brunswick or Gettysburg formation. At one locality near Clinton, New Jersey, they occur in the Lockatong and Stockton formations. South of Reading, Pennsylvania, the largest area of conglomerate in the basin extends across the Gettysburg formation into the New Oxford formation. Most of the gravels in the conglomerates and breccias are derived from lower Paleozoic formations, although some pebbles indicate derivation from sources as high as Devonian and as low as pre-Cambrian, and in one instance Triassic basalt forms boulders and cobbles in a fanglomerate. The chief lithologic constituents among the gravels are quartzite and sandstone pebbles, generally subrounded to subangular, and limestone or dolomite pebbles, generally subangular to angular.

In 1913 Lawson proposed the term "fanglomerate" for cemented deposits of angular to subangular pebbles and cobbles which were formed in alluvial fans. The term was proposed to replace "breccia" which had too wide a connotation, and "conglomerate" which denotes an aggregate of well-rounded pebbles. As he stated, it is "... a descriptive term with specific connotation as to both the character of the rock and its mode of formation" (1913, p. 330). The border conglomerates and breccias of the Newark series were suggested as possible examples of fanglomerates. Inasmuch as most of these rocks probably originally accumulated in fans at the base of the western border fault scarp and contain very few well-rounded pebbles, they are referred to as fanglomerates in this paper, but they are differentiated as conglomerate, where locally composed of moderately rounded pebbles and cobbles; and as breccia, where locally composed of angular fragments.

BORDER BRECCIA DEPOSITS

The sedimentary breccia deposits generally consist of subangular to angular limestone and dolomite fragments in a red, sandy matrix. One of these deposits, the Montville breccia (Kümmel, 1898, p. 53–55), consists mainly of boulders and cobbles of granite and gneiss and minor amounts of basalt, quartzite, sandstone, and limestone fragments. In addition to occurring as separate deposits, breccia is found in the border conglomerates as angular facies nearest the border fault, and as beds of breccia intercalated with beds of more rounded gravels. Most of the breccia deposits are of the type described by Norton (1917, p. 167–169) as bajada breccia, although there are some breccia deposits formed through accumulations of talus creep and rockfalls. An outcrop in a typical limestone breccia deposit is shown (Pl. 2, fig. 3). Kümmel has given excellent descriptions of these breccias (1897, p. 52–54; 1898, p. 53).

QUARTZITE CONGLOMERATES

General considerations.—The conglomerates of the northwestern border of the Triassic basin have been named by Kümmel the "quartzite conglomerates" (1897, p. 50). Their most abundant pebble types are quartzite and sandstone, although limestone, dolomite, gneiss, shale, slate, or quartz pebbles are present in minor amounts. Although the quartzite and sandstone pebbles are more resistant to abrasion than the limestone and dolomite pebbles, they are generally more rounded.

The largest area of conglomerate in the basin is south of Reading, Pennsylvania. This deposit has been studied in detail by McLaughlin (1939, p. 59-74), who has named it the "Robeson conglomerate" and has correlated it with the Brunswick formation of New Jersey and New York. The deposit is about 40 miles long and about 8 miles wide. The second largest area between the Susquehanna River and the Hudson River is at Pattenburg, New Jersey. It extends for approximately 5.5 miles along the border fault and 3.5 miles out into the basin. The other deposits of conglomerate along the border fault range down in outcrop area to less than a mile in greatest diameter.

Although confined mainly to a fringe along the western border fault, the conglomerates in places extend completely through the series and even across the basin. The Robeson conglomerate is such an example, and Kümmel (1898, p. 46) has described conglomerate beds at points well across the basin in its northern part at Hackensack, Maywood, Rochelle Park, and Paterson, New Jersey. Conglomerate beds grade laterally into sandstone and shale. Interbedding of conglomerates with breccia, sandstone, and shale is common.

Plate 2 shows two typical exposures of Triassic conglomerate in the Triassic basin.

Dimensions of pebbles.—The average diameter of the pebbles in the border conglomerates is approximately 3 inches, as estimated at 16 deposits. The average diameter of the largest boulders in 31 outcrops in 17 conglomerate deposits is approximately 18 inches. At 20 of these outcrops the largest boulders were quartzite or sandstone and ranged from 5 to 34 inches, averaging 16 inches in greatest diameter. At 5 outcrops limestone or dolomite boulders were the largest; these ranged from 12 to 44 inches in diameter, with an average diameter of 23 inches. At one outcrop (in the southeastern New Oxford phase of the Robeson conglomerate) a quartz boulder 8 inches in greatest diameter was the largest in an extensive outcrop.

Degree of rounding.—Degree of rounding of pebbles and cobbles in the conglomerates was determined by visual estimate based on Tester's evaluation of rounding (1931, p. 3–11). At every conglomerate area hundreds of cobbles and pebbles were examined for well-rounded specimens showing no trace of original shape. This method permitted rough estimate of degree of rounding of many samples, rather than precise analysis of necessarily limited numbers of samples, and proved advantageous since many conglomerate zones exhibited considerable variation in rounding of pebbles both throughout the area of outcrop and from one bed to another. The size of the pebbles and cobbles did not permit collection of hundreds of samples from each area for laboratory analysis, and there was insufficient time for precise mechanical analysis in the field because of the many square miles of Triassic conglomerate between the northeast end of the Carlisle Prong and the Hudson River.

FIGURE 1. TRIASSIC CONGLOMERATE
South of Ladentown, N. Y.

FIGURE 2. TRIASSIC CONGLOMERATE AT "PEBBLE BLUFFS"
On the Delaware River

FIGURE 3. TRIASSIC LIMESTONE BRECCIA
One mile northwest of Stonerville, Penn.

FIGURE 4. CAMBRIAN QUARTZITE GRAVEL ON THE BANKS OF OCONALUFTEE RIVER, GREAT SMOKY MOUNTAINS.
Fountain pen gives scale.

TYPICAL EXPOSURES OF TRIASSIC FANGLOMERATE OF THE
NEWARK BASIN, AND A MODERN STREAM DEPOSIT IN THE GREAT SMOKIES.

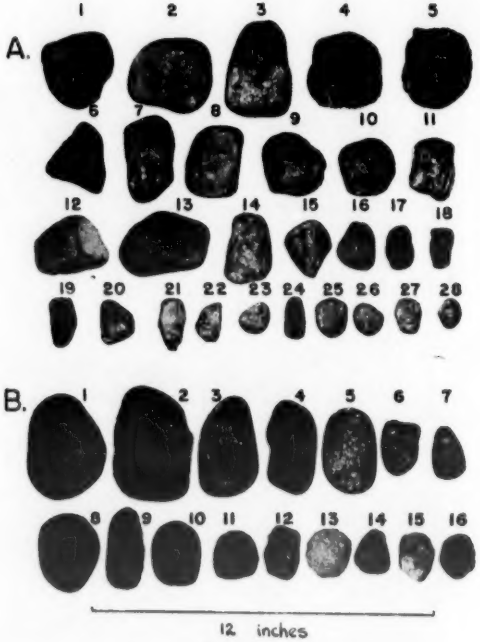

FIGURE 1.

(A) PEBBLES FROM OUTCROP ON MOHNTON-GOUGLERSVILLE, PENN., ROAD.
All pebbles are quartzite.

(B) PEBBLES FROM OUTCROP ON SOUTHEAST SIDE OF UNION HILL, NEAR SUFFERN, N. Y.
All pebbles are quartzite and quartzitic sandstone. Foot rule divided in inches gives scale.

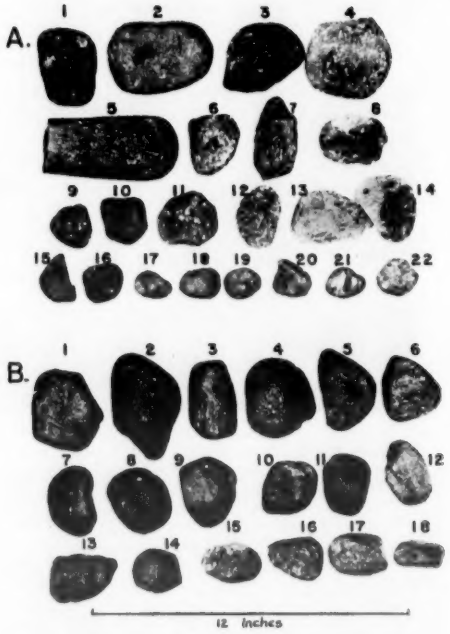

FIGURE 2.

(A) PEBBLES FROM CONGLOMERATE. IN ROAD CUT ONE MILE EAST OF LISBURN, PENN.
Pebbles 1 to 11 and 15 to 20 inclusive are quartzite and quartzitic sandstone. Pebbles 12 to 14 are limestone, and pebbles 21, 22 are quartz.

(B) PEBBLES FROM ROAD CUT 2 MILES EAST OF LITTLE YORK, N. J.
All pebbles are quartzite and quartzitic sandstone. Foot rule divided in inches gives scale.

TYPICAL PEBBLES FROM FOUR TRIASSIC BORDER CONGLOMERATES.

It was found that there is little variation among the conglomerates in the degree of rounding of their gravels. The largest proportion of their pebbles are of curvilinear forms with subrounded and subangular shapes somewhat less abundant. Angular forms are fairly common, and rounded forms are by far the least abundant. These relations are illustrated by Plate 3, photographs of quartered samples from four typical border conglomerates in the area studied. In Figure 1 (A) of Plate 3, application of Tester's method of roundness determination discloses that 17 of the pebbles are curvilinear in shape, 8 subrounded, and 3 subangular. In Figure 1 (B) one pebble is rounded, 13 are subrounded, and 2 are curvilinear. In Figure 2 (A) 4 pebbles are subrounded and 15 curvilinear; and in Figure 2 (B) 7 pebbles are subrounded, 10 are curvilinear, and one is subangular.

Determination of source beds.—All but one of the writers who have attempted to identify the origin of the Triassic border sediments have assigned to the gravels sources ranging in age from pre-Cambrian to Devonian. This has been summarized in Table 1. The table also contains a number of source identifications of Triassic gravels which have been reported to the writer by personal communication.

In additon to the gravels of Paleozoic derivation some crystalline pebbles have been found in Triassic border sediments at the following areas: (1) a gneiss fragment found in fanglomerate south of Stony Point, New York, by A. N. Strahler, (2) a pre-Cambrian cobble found by Wheeler and Miller in a road cut 4 miles southwest of Suffern, New York, (3) granitic and gneissic cobbles and boulders in a fanglomerate from 3 to 7 miles southeast of Pompton Lakes, New Jersey, near Montville, New Jersey, (4) crystalline fragments as a minor constituent in a limestone breccia deposit on the west side of the Delaware River, within half a mile of the fault line, (5) one granite pebble in the Robeson conglomerate south of Reading, Pennsylvania, in an outcrop at the intersection of the roads to Grill and the County Farm. The source of the crystalline pebbles and cobbles is obviously the pre-Cambrian basement of the Reading Prong.

Rounded to subrounded quartz pebbles are common minor constituents in nearly all the fanglomerates along the highland border of the Triassic basin. These may be derived from quartz veins in the Reading Prong and in the Paleozoics, or they may be reworked pebbles from Paleozoic pebbly quartzites. A third possibility is that they represent material transported northwestward in pre-Newark time from the old highland mass of Appalachia and spread out over the Paleozoic terrain northwest of the present location of the Triassic basin. During Newark sedimentation streams consequent upon the highland border fault scarp would have carried them back into the basin along with material eroded from the Paleozoic terrain.

DISTANCE OF TRANSPORTATION OF TRIASSIC BORDER CONGLOMERATES

General statement.—Direct evidence for major river deposition of any of the western border conglomerates is lacking. The total area of outcrop of the conglomerate deposits might be an index of the size of the stream. The Robeson conglomerate is by far the largest area of conglomerate in the basin, yet its great size might well indicate high, massive mountains bordering it on the north (McLaughlin, 1939, p. 73, 74). The bedding of the conglomerates is generally poor; torrential bedding is

TABLE 1.—*Age assignments for source beds of Triassic gravels in the Newark basin*

Writer	Area Studied	Description of conglomerate	Age assignment of source rocks
Kümmel (1898, p. 46–47)	Hackensack, N. J., to Paterson, N. J., conglomerate beds of Brunswick age	Sandstone and quartzite gravels	Silurian Green Pond cgl.
Darton and Kümmel (Darton, Bayley, Salisbury, and Kümmel, 1908, p. 8)	Passaic quadrangle, N. J., Montville and Pompton Lakes fanglomerates	Sandstone and quartzite gravels	Silurian Green Pond cgl.
Wherry (1913, p. 120)	South of Spring Valley, Penna., fanglomerate outcrops	Quartzite gravels..... Limestone gravels....	Silurian Shawangunk Cambro-Ordovician limestone
Stose and Bascom (1929, p. 10)	Fairfield and Gettysburg quadrangles, Penna., Arendtsville fanglomerate	Aporhyolite gravels... Quartzite and sandstone gravels.....	Pre-Cambrian Cambrian
Bascom, Wherry, Stose, and Jonas (1931, p. 47–48)	Quakertown-Doylestown District, Penna., border fanglomerates	Quartzite gravels..... Limestone gravels...	Devonian quartzite near-by Paleozoic limestones
Meyerhoff and Olmsted (1936, p. 38)	Newark basin fanglomerates	Quartz and quartzite gravels..... Limestone gravels....	Middle and upper Paleozoic formations Cambro-Ordovician limestone
Bascom and Stose (1938, p. 102)	Honeybrook and Phoenixville quadrangles, Penna., conglomerates south of Reading, Penna.	Quartzite gravels..... Limestone gravels....	Silurian quartzite adjacent Paleozoic limestones
McLaughlin (1939, p. 71–74)	Robeson conglomerate south of Reading, Penna.	Quartzite gravels	Cambrian quartzite
Miller (1939a, p. 273)	South of Spring Valley, Penna., fanglomerate outcrops	Quartzite gravels..... Limestone gravels....	Silurian Shawangunk or Green Pond cgl. Cambro-Ordovician limestone (made with reservations based on magnesia content of gravels)

TABLE 1.—(*continued*)

Writer	Area Studied	Description of conglomerate	Age assignment of source rocks
Stose and Jonas (1939b, p. 117)	York County, Penna., banks of Yellow Breeches Creek, fanglomerate outcrop	Sandstone boulder containing brachiopod	Suggested Devonian Oriskany age
(1939b, p. 116)	Same area	Reported discovery by Bissell of sandstone boulder containing brachiopod identified as *Meristella lata*	Devonian Oriskany
Miller, R. L. (personal communication)	Delaware River Valley conglomerate outcrops	Quartzite gravels.....	Silurian Tuscarora
		Limestone gravels....	Cambro-Ordovician Kittatinny
	Robeson conglomerate south of Reading, Penna.	Quartzite gravels.....	Silurian Tuscarora
		Limestone gravels....	Cambro-Ordovician Kittatinny
		One pebble.....	Devonian Oriskany or Esopus
Wheeler, G., and Miller, R. L. (personal communication)	Darlington, N. J., conglomerate outcrop	Limestone cobble containing brachiopods	Devonian Helderbergian
Johnson, Meredith (personal communication)	Triassic conglomerates of New Jersey	Quartzite gravels	Silurian Shawangunk or Devonian Skunnemunk
		Limestone gravels	Mostly from Cambro-Ordovician Kittatinny
Anderson, R. J. (personal communication)	Rockland County, N. Y., fanglomerates	Fossiliferous boulders	Lower Ordovician to lower (?) Devonian
Meyerhoff, H. A. (personal communication)	Newark basin fanglomerates	Unspecified lithologies	As young as Mississippian Pocono

common (Pl. 2, figs. 1, 2). Their primary structures suggest deposition by short steep streams, rather than by large master streams. The size of the gravels in the conglomerates is no index to the size of the stream which deposited them. With the exception of the primary structures of the deposits, only the degree of rounding of the gravels and their determinable source beds provide clues to their distance of transportation.

Significance of degree of rounding.—Kümmel (1898, p. 52) stated in regard to the Triassic quartzite conglomerates: "All the constituent materials are well rounded, a fact which in the case of the hard quartzite indicates a long period of attrition."

When examined and measured by the Tester method of roundness analysis, how-

ever, the conglomerate pebbles are by no means all well rounded. Furthermore, more recent work has shown that degree of rounding of pebbles is a not infallible index to the total distance of transportation.

Marshall (1927, p. 507–532) has described three types of wear and reduction of rock fragments. They are "grinding," or the crushing of smaller fragments by larger fragments; "impact," or definite blows dealt by one fragment to another; and "abrasion," or rubbing between fragments.

According to R. Dana Russell (1939, p. 39):

> "Though 'abrasion' (rubbing) usually results in rounding of the particles, 'grinding' apparently always produces angular small grains, though the large 'grinders' may become rounded. Theoretically 'impact' could result either in rounding or angularity, depending upon the relationship between the energy of impact and the sizes of the grains involved. Whether a grain becomes rounded or angular would appear to be a function of the sizes of the chips and fragments broken off in comparison with the size of the particle as a whole. If the fragments broken off are relatively large, the particle would remain angular or become more so; if they are small in comparison with the size of the grain as a whole, the general effect would be an increase in roundness."

Field evidence for the effect of "impact" in maintaining or increasing angularity of particles during transportation is given in the studies of Fugger and Kastner (1895, p. 143–144), who state that in the Salzach River, a steep mountain stream of the Austrian Alps, pebbles remain angular through long transportation, since their loss in size during movement from the river's source to the plains below is principally caused by impact and shattering of the pebbles.

Under still other favorable conditions, rounding of pebbles during transportation is inhibited. Oldham (1894, p. 464–465) described the gravels in alluvial fans of western and central Asia as poorly rounded in the outermost limits of fans many miles from their source. He accounted for this slow rate of rounding by suggesting that

> "... when the streams flow after rain they are generally so loaded with debris as to be rather of the nature of fluid mud than water, and in this the fragments of rock seem to be carried along *en masse* without being worn against each other to the same extent as in a mountain torrent."

In 1940 Krumbein (p. 639–676) reported on a detailed study of flood-deposited gravels of the San Gabriel Canyon in the San Gabriel Mountains of southern California. In his Plate 1 he shows granitic and dioritic pebbles from two stations in the canyon. At one station the pebbles are angular, and at the other station 5½ miles downstream they are subangular according to the Tester method of roundness determination, or of mean roundness .44 as calculated by Krumbein's method. According to Krumbein the rate of rounding of the pebbles should have been higher than normal. He states, "In contrast to normal streams, the large volumes, high velocities, and enormous loads of streams in flood may be expected to accelerate such processes as abrasion and rounding. . . ." (p. 642). The large volumes and high velocity of flood conditions in the San Gabriel Canyon may, however, result in the "*en masse*" form of transportation described by Oldham which would tend to reduce the rate of gravel rounding rather than increase it.

While the method of transportation described by Fugger and Kastner differs greatly from that described by Oldham, they both result in maintenance of the angular shapes through long transportation. Other studies of pebble rounding point

to progressive rounding of pebbles, cobbles, and boulders with increasing distance of transportation.

In his study of the alluvial deposits of the Kis-Szamos River in Transylvania Von Szádeczky-Kardoss (1932–1933) found that granite, gneiss, andesite, and quartzite fragments under 32 mm. in diameter remain angular over relatively long distances of travel, while fragments over 32 mm. in diameter show a very rapid rate of rounding in the beginning (1933, p. 250). He shows that in about 7 miles of transportation quartzite reached a rounding value approximately midway between complete angularity and complete rounding (1933, p. 259).

In the Arroyo Seco of Los Angeles County, California, Krumbein (1942, p. 1384–1386) found that in a little over 10 miles of flood transportation the mean roundness of granodiorite pebbles increased only from .20 to .38, about half the rate of rounding found at San Gabriel Canyon. In the San Gabriel Canyon Krumbein measured rock fragments roughly 32 to 64 mm. in diameter, whereas in the Arroyo Seco he studied rock fragments 16 to 32 mm. in diameter. The observations of Szádeczky-Kardoss explain the discrepancy in rates of rounding.

Wentworth's study (1922, p. 103–114) of the rate of rounding of pebbles in the Russell Fork of Big Sandy River in southwestern Virginia deals exclusively with yellow vitreous to white sugary pre-Carboniferous quartzite. His Figure 42 (1922, p. 111) shows a very rapid rate of rounding in the first 2 miles of transportation, during which the degree of rounding (expressed as roundness ratio, or $\frac{r_1}{R}$ where r_1 is the radius of curvature of the sharpest developed edge, and R is the mean radius of the pebble) changed from 0 to .108. In the following 34 miles of transportation the roundness ratio increased to only .16. A profile of the river is given in his Figure 37 (1922, p. 105), and in the first 2 miles of transportation the river falls at the rate of 350 feet per mile; in the next 25 miles the drop averages only 20 feet per mile.

A distinct break in the curve of his graph, Figure 42, showing effect of transportation on rounding, occurs at the end of the second mile of transportation. Since the steepest portion of this stream is within the first 2 miles of its course, this would indicate that the higher velocities in the steeper reaches result in more violent and more frequent interfragment contacts. In a stream in which impact is not so violent as to produce a significant amount of shattering of the transported fragments, and there is no evidence of "*en masse*" transportation as described by Oldham (1894) it is suggested that stream declivity and velocity are controlling factors in the rate of rounding.

Bonney (1888, p. 54–61) elucidated this principle. He compared the rate of rounding of rock fragments in Alpine glacial torrents with the rate of rounding of rock fragments in tributaries of the Po River which had flowed over 35 to 50 miles of plain. He concluded from this study that "...fairly well rounded pebbles of a rock with hardness not exceeding 6 signify at least either a rapid descent of three thousand feet or a journey at a less speed of sixty miles" (1888, p. 60).

In the course of field study of pebble rounding in streams of the Appalachian highlands, Sherman K. Neuschel studied the rates of pebble rounding on Russell Fork of Big Sandy River, the locale of Wentworth's earlier work. After examining

the photographs in Plates 2 and 3 of the present paper and other suites of pebbles collected by the writer from the Triassic conglomerates, he stated (personal communication) that the degree of rounding of these quartzite pebbles from the Triassic conglomerates was present in Russell Fork after 5 miles of transportation.

Figure 4 of Plate 2 shows Cambrian quartzite gravel in the bed of Oconaluftee River, 1.2 miles northwest of Smokemont, North Carolina, in the Great Smokies National Park. Cambrian quartzite outcrops throughout the watershed of the Oconaluftee River in North Carolina. Although the greatest possible distance of transport of these pebbles is 7.2 miles (measured to the headwater divide of the main branch of the river), most of these pebbles have probably traveled less than that. The degree of rounding of the gravel is virtually identical with the degree of rounding of gravels in the Triassic border conglomerates.

Thus the degree of rounding of the pebbles in the Triassic border conglomerates could have been produced by short steep streams heading in a moderately high mountain mass along the western border of the Triassic basin, or by long rivers heading many miles back in the Triassic Appalachians. The studies of Bonney, Wentworth, Szádeczky-Kardoss, and Neuschel have demonstrated that neither agency can be appealed to as the only means by which the Triassic quartzite conglomerates could have received their observed degree of rounding. The degree of rounding of these pebbles does not prove long-distance major river transportation.

Significance of source beds of conglomerate pebbles.—Distance of transportation can be determined by lithologic character and fossil content of pebbles, provided one can fix with reasonable accuracy the nearest available source at time of deposition. At present the older formations are nearer the Triassic basin, younger formations are more remote. The youngest Paleozoic pebbles positively identified are: (1) Devonian Oriskany in conglomerate near the Susquehanna River; (2) Devonian Oriskany or Esopus in the Robeson conglomerate. These beds now outcrop within 20 miles of the Triassic basin. (3) Devonian Helderberg limestone in conglomerate near Suffern, New York. The Helderberg now outcrops within 10 to 15 miles of the Triassic basin. (4) Upper Devonian Skunnemunk pebbles possibly present in the New Jersey Triassic. The Skunnemunk conglomerate now outcrops within 10 miles of the Triassic basin.

The Oriskany beds could have outcropped even closer to the Triassic basin during Newark time than they do at present, conceivably even in the scarp of the border fault. Consequently the figures given above are maximum distances of transportation with minimum distances possibly far less. A Cambrian to Devonian sequence of beds now outcrops in the Reading Prong. Although no Helderberg limestone has been found in the New Jersey section of the Reading Prong highlands, Kümmel (Lewis and Kümmel, 1940, p. 93–94) has stated that a thin representative of the Helderberg may occur there since a continuous section is nowhere exposed. Helderberg limestone does occur in the Reading Prong in Orange County, New York, near Bellvale, Highland Mills, and Cornwall. At Cornwall, which is within 15 miles of the Triassic basin and 25 miles of where Helderberg limestone was found in the Triassic by Wheeler and Miller, Ries (1897, p. 426) has reported 50 feet of Helderberg limestone. The Skunnemunk conglomerate now outcrops within 10 miles of

the Triassic basin and could have outcropped even closer to the basin in Triassic time. The lithologic character and fossil content of pebbles in the Triassic conglomerate indicate transportation for maximum distances of 15 to 25 miles. Inasmuch as all of the Paleozoic formations discussed above, except possibly the Helderberg limestone, could have outcropped in the Triassic fault scarp, the minimum distance of transportation as indicated by the lithologic character and fossil content of the Triassic pebbles could be considerably less.

Conclusions.—The positively identified source beds of gravels in the Triassic border fanglomerates fail to prove transportation for greater distances than 25 miles, and, since these beds should have outcropped closer to the basin during the Triassic, they suggest even shorter distances of transportation. The degree of rounding of the gravels could have been produced by long master streams or by steep streams less than 10 miles long. The primary structures of the gravels strongly favor the hypothesis of transportation by short, steep streams rather than master streams as held by Meyerhoff and Olmsted.

RELATION OF APPALACHIAN DRAINAGE TO THE HIGHLAND BORDER SEDIMENTS OF THE NEWARK SERIES

MAJOR APPALACHIAN RIVERS

Hudson River.—The Hudson River enters the northern end of the Triassic basin. Triassic sediments occur along the west side of the river in front of a fault-line scarp which continues east of the river as an important topographic feature. Although no Triassic has been found east of the river, there the scarp faces an area of low relief that has been interpreted as a pre-Triassic peneplane from which Triassic sediments have been stripped (Johnson, 1933, p. 87). These relations suggest that Triassic sediments were once present on both sides of the Hudson at its point of exit from the bordering highlands.

Where the Hudson leaves the Highlands a breccia deposit on the west bank of the river consists mainly of angular limestone fragments, with approximately 2 per cent angular quartzite fragments in a red sandy silt matrix. Southward, along the river bank, the breccia grades very rapidly into shale with interbedded sandstone. The angularity of the breccia fragments and the predominance of limestone in the breccia indicate deposition by a short, steep stream. A large area of well-rounded Triassic gravels derived from the upper Hudson Valley which might suggest the presence of an ancestral Hudson is notably absent.

Delaware River.—Where the Delaware River enters the Triassic basin it is flanked on the east side by two small areas of quartzite border conglomerate separated by a small area of limestone breccia. West of the river a third zone of quartzite conglomerate outcrops. None of the three areas extends out into the basin more than $1\frac{1}{2}$ miles and none is more than 1.2 miles wide. Where the river enters the basin it cuts through a breccia deposit containing subangular to angular Cambro-Ordovician limestone fragments.

Fresh outcrops in the conglomerate area nearest the river on the eastern bank are present only along the road from Little York to the river. A 30-foot section has

white to gray brecciated quartzite cobbles and boulders in a tan clay matrix; the cobbles and boulders are badly crushed by movement along the adjacent border fault. Between this conglomerate area and the next to the east lies a small area of Cambro-Ordovician limestone breccia. The next conglomerate to the east outcrops in high, unweathered road cuts along the river road on the east bank of the Delaware, 1½ miles north of Milford, providing the best section of Triassic border conglomerate exposed northeast of the Susquehanna River, possibly the best in the whole basin. (*See* Figure 2 of Plate 2.) These outcrops were named the "pebble bluffs" by Kümmel (1898, p. 53). The section consists of coarse conglomerate beds striking N. 30° E. with dips ranging from 15° to 23° NW.

Pebble counts were made in four exposures in the "pebble bluffs." At the southeastern end of the deposit 45 per cent of the pebbles are gray Tuscarora quartzite, 40 per cent reddish Tuscarora quartzite, 6 per cent Kittatinny limestone, 5 per cent sandstone, and 4 per cent quartz. Where the Triassic beds are gray the quartzite pebbles are gray and unaltered; where red, the quartzite and sandstone pebbles are prevailingly red. Some of the red quartzite pebbles have gray cores and red jackets. The southeasternmost exposures in the "pebble bluffs" have practically no limestone pebbles and a considerable percentage of interbedded sandstone and sandy shale. To the northwest, or toward the fault, the percentage of limestone pebbles increases up to 20 per cent, their size and angularity increase, and the pebble beds make up an increasing amount of the section until at least 60 to 75 per cent of the section becomes pebble bearing. At the northwesternmost good outcrop no bedding is visible through about 40 feet of outcrop. The pebbles are unsorted, unbedded, and prevailingly subangular with many angular pebbles. Limestone pebbles constitute about 20 per cent.

The average degree of rounding of the pebbles throughout the deposit ranges from angular to subrounded; the most common shapes are subangular and curvilinear. Along the "pebble bluffs", a distance of less than a mile, one can trace downward through the section a progressive increase in rounding, from dominantly subangular and angular shapes in the northwestern part to curvilinear and subrounded shapes in the southeasternmost outcrop.

Triassic outcrops along the west bank of the river reveal beds of breccia containing about 90 per cent angular blocks of Cambro-Ordovician limestone up to about 5 inches in diameter and about 10 per cent fragments of angular gneiss. The beds of breccia are interbedded with greater thicknesses of sandstone and shale and persist for approximately .6 mile southeast of the fault line. Beyond, the Triassic is entirely sandstone and shale. Adjacent to the fault line on the east bank of the river are outcrops of breccia composed almost entirely of limestone fragments.

The small conglomerate deposit just west of the river lacks fresh outcrops, but mantling the surface of the small hill which it forms are many subangular to curvilinear pebbles, cobbles, and boulders of quartzite.

Summary.—Where it enters the basin the Delaware flows through a breccia deposit containing limestone and some gneiss fragments. Areas of border conglomerate flank the breccia and differ from other conglomerate areas along the border fault only in their somewhat small area of outcrop. The degree of rounding of the pebbles

and cobbles in the conglomerates is no higher than in other conglomerates, and the source beds indicated by lithologies are Cambro-Ordovician and Silurian. There is no suggestion that Triassic border sediments near the Delaware were laid down by a large stream, such as an ancestral Delaware River, nor that they are significantly different from conglomerates along other parts of the Triassic border.

Lehigh River.—The Lehigh River does not enter the Triassic basin at the present time. Its southeast trend changes at Allentown to a northeast course toward the Delaware. Meyerhoff and Olmsted (1936, p. 38) apparently referred to the Lehigh when they wrote: "One large deposit of quartzite conglomerate situated south of Spring Valley, Pennsylvania, is not associated with a modern river, but its position south of the broad gap at Bethlehem leaves no doubt as to its origin and initial fluvial connections."

Mackin (1938, p. 39) has pointed out that the presence of this broad gap at Bethlehem does not prove that it was once occupied by the Lehigh. The Bethlehem gap coincides with the outcrop of Cambro-Ordovician limestone that consistently forms lowlands throughout the length of the Appalachians and would be present if there were no major river within many miles of the area. He also remarked that the lithologic character of the Spring Valley conglomerate does not prove that the material was deposited by the Lehigh. The conglomerate, while possessing some subrounded to subangular quartzite pebbles, also contains angular quartzite boulders up to 3 feet in diameter. In a ravine trenching the north face of the hill formed by the conglomerate are beds filled with angular limestone and quartzite fragments. The extreme angularity of this material indicates a very limited distance of transportation. The deposit is no larger than a number of other areas of border conglomerate not associated with large streams and could reasonably be interpreted as the deposit of a relatively small stream heading a short distance back of the Triassic border fault scarp.

According to Wherry (1913, p. 120) the quartzite pebbles in this deposit came from the Silurian Shawangunk quartzite or the Silurian Green Pond conglomerate which now outcrops in northern New Jersey. The limestone pebbles, he observed, were derived from Cambro-Ordovician limestones of the Great Valley. B. L. Miller (1939a, p. 273) has stated that the quartzite pebbles in the deposit were probably derived from the Shawangunk quartzite and the limestone pebbles from Cambro-Ordovician limestones. He noted, however, that the low magnesia of the limestone pebbles did not agree with such a correlation.

There are four other gap systems, besides the Bethlehem-Spring Valley system, through which the Lehigh may have flowed before its capture by the Delaware River. Four of the five would lead the Lehigh across Triassic border conglomerates. Examination of the gap systems on the topographic maps of the Allentown, Slatington, Quakertown, and Boyertown quadrangles and in the field, however, shows that the systems are mere sags in the ridge crests of the general upland of the Reading Prong. Such sags may be the result of local stream erosion, and weathering in some cases partly controlled by strips of infaulted Paleozoics. If only one clearly marked gap system trended across resistant rocks in this area, and if this system connected with the greatest area of quartzite conglomerate within miles of the region, the case for

genetic correlation would be good. Instead there are at least five hypothetical gaps across the Reading Prong, and to match these there is a nearly continuous strip of Triassic conglomerate along the border fault.

Schuylkill.—The Schuylkill enters the Triassic basin opposite the largest area of quartzite conglomerate in the basin, which has been termed the "Robeson conglomerate" by McLaughlin (1939, p. 60). He has traced the conglomerate 27 miles west of the Schuylkill and into Furnace Ridge, south of Cornwall. Another tongue of the Robeson conglomerate extends along the southern base of South Mountain, west of Reading. The end of the Furnace Ridge tongue is about 40 miles west of the eastern termination of the conglomerate area. Any major river originating in post-Triassic time in this general area would have had a 40-mile east-west zone in which it could have taken its course and still flowed across the zone of conglomerate. Disregarding the western fringes of the Robeson conglomerate, the main body extends from its eastern end 25 miles in a north of west direction to the vicinity of Reinholds. A post-Triassic Schuylkill would thus have had a 25-mile zone in which to take its course and be opposite the main body of conglomerate. Thus the fact that the Schuylkill enters the Triassic basin within the limits of the Robeson conglomerate cannot be considered extraordinary.

The main body of the Robeson conglomerate lies within the confines of a great ring-shaped dike of Triassic diabase. North of the dike the Triassic consists of about 1900 feet of sandstone and shale with Cambro-Ordovician limestone-dolomite breccia in the uppermost beds nearest the fault line. In one outcrop near Grill a pre-Cambrian crystalline pebble was found.

The largest boulders and heaviest concentration of conglomerate were not observed along the northern border of the conglomerate but about $1\frac{1}{2}$ miles south of its northern border at the following locations: (1) a mile south of Mt. Penn School, where curvilinear quartzite boulders up to 8 inches in diameter occur; (2) half a mile southwest of Angelica, where curvilinear to subrounded 4- and 6-inch quartzite cobbles are common and quartzite boulders up to 10 inches in diameter occur; and (3) a mile east of Gouglersville, where nearly the whole visible Triassic sequence is pebble bearing.

Quartzite and sandstone pebbles, cobbles, and boulders form the principal constituents of pebble beds in the conglomerate throughout the northern and central parts. In the northern section Cambro-Ordovician limestone makes up some 10 to 20 per cent of the pebbles, but in the central section quartzite and sandstone pebbles are the more prevalent constituents.

In the southeast and southern part of the conglomerate subrounded to rounded pebbles and cobbles of quartzite and quartz in equal amounts form the New Oxford phase of the conglomerate. All constituents were apparently derived from a crystalline terrain with infolded quartzite ridges to the south.

Many outcrops of the Robeson conglomerate contain pebbles of vein quartz. Throughout most of the outcrops they rarely make up more than 5 per cent of the pebbles and are generally subrounded. In the field discussion Meyerhoff stated that he considered the quartz pebbles in conglomerate near the northern margin of the basin to be reworked from the upper Paleozoic pebbly quartzites. An

alternative possibility is that they represent material transported northward in pre-Newark time from the old highland mass of Appalachia and spread out over the Paleozoic terrain north of the present location of the Triassic basin. During Newark sedimentation streams consequent upon the highland border fault scarp would have carried them back and into the basin along with material eroded from the Paleozoic terrain.

Cambrian quartzite pebbles are the oldest Paleozoic constituents in the Gettysburg phase of the Robeson conglomerate. Devonian Oriskany or Esopus is the youngest. (*See* section on Nature of the Triassic Sediments, Determination of Source Beds.) The great bulk of the gravels are Silurian quartzite and sandstone, and the second most abundant constituent Cambrian and Ordovidian limestone fragments. McLaughlin has stated that the Robeson conglomerate quartzite pebbles were entirely derived from Cambrian quartzite (1939, p. 71–74).

Mackin (1938, p. 38) points out that dolomite and limestone breccia at least 1000 feet thick outcrops where the Schuylkill enters the Triassic basin. The breccia consists of angular dolomite and limestone fragments in a red silt matrix. The predominance of limestone and dolomite fragments in the deposit, the lack of bedding, and the angularity of the material indicate deposition by a short steep stream heading in a high scarp to the north. To the south along the Schuylkill the breccia grades abruptly into sandstone and shale. Mackin concludes that the Schuylkill could not have been near its present position at the time of deposition of the breccia and shales and that

> "local conditions of sedimentation varied widely from time to time during the Triassic, and . . . no single Triassic stream maintained a course in any way comparable with that of the Schuylkill throughout the deposition of the Newark series" (p. 38).

Susquehanna River.—Triassic sandstone and shale extend close to the northern boundary fault along both sides of the Susquehanna River valley. Immediately adjacent to the fault on the east side of the river Triassic sandstone is intercalated with beds containing fragments of Paleozoic shale and a small number of angular limestone, quartzite, and sandstone fragments. On the west side of the river the fault line is bordered by a narrow fringe of Triassic breccia which consists mainly of angular fragments of limestone in a sandy matrix. There is no area of conglomerate opposite the point where the Susquehanna enters the Triassic basin. Breccia, sandstone, and shale are the only rocks represented in these Triassic border sediments.

An area of conglomerate extends from about 2 to 6½ miles east of the Susquehanna River; its center is opposite the point where Swatara Creek enters the basin. This conglomerate coarsens eastward; the coarsest phase is about 5 miles east of the river where quartzite and sandstone pebbles are the principle constituents.

The breccia at the river was traced westward to about 2½ miles west of the river. Two miles west of the river it has angular fragments of shale, limestone, and quartzite.

From about 4 to 6 miles west of the Susquehanna River is a fanglomerate deposit here referred to as the Lisburn fanglomerate. The town of Lisburn is on its western margin. The area is about 2 miles wide parallel to the boundary fault and extends some 4 miles out into the basin. In its upper part, near the fault line, the deposit consists mainly of angular fragments of limestone, quartzite, and sandstone, in a

sandy matrix. At one outcrop many small Paleozoic shale fragments are incorporated in the breccia. In the southern part of the deposit the pebbles are rounder. About a mile east of Lisburn the exposed conglomerate has subangular to subrounded quartzite, sandstone, and limestone pebbles. One cobble of Devonian Oriskany sandstone was found. Much of the quartzite and sandstone gravels are of typical Silurian lithologic character.

No breccia or conglomerate is known adjacent to the boundary fault between the Lisburn area and the northeast end of the Carlisle Prong. Except for a breccia deposit fringing the border fault, the Triassic sediments between the Lisburn conglomerate and the Susquehanna River are mainly sandstone with interbedded shale. Intercalated in the Triassic sandstone throughout this area are occasional pebble beds which appear to be tongues heading in the main mass of conglomerate to the west. The number and thickness of the pebble beds decrease to the east or toward the river. Four miles east of Lisburn the pebble beds are so rare that the sequence is not truly conglomeratic. In a road cut on the New Cumberland-York highway where it ascends from the Susquehanna flood plain to the Triassic upland a thick section of Triassic sandstone contains a few pebble beds of two types: quartzite and limestone pebbles in a sandy matrix, and green Paleozoic shale fragments with some quartzite and limestone pebbles. The shale fragments are prevailingly angular, the limestone fragments mainly subangular, and the quartzite pebbles subangular to subrounded.

Conclusions.—The Hudson River is associated with Triassic breccia deposits, sandstone, and shale. The Delaware, a large major river, is associated with relatively insignificant deposits of conglomerate and breccia. Any one of five courses may be postulated for entry of the Lehigh River into the basin, making correlation a highly subjective procedure. The Schuylkill, far smaller than the Delaware, is associated with conglomerate many miles in length and breadth; and the Triassic sediments nearest the largest river of all, the Susquehanna, are breccia deposits and a few pebble beds in sandstone and shale. There is little or no correlation between present courses of major Appalachian rivers and Triassic conglomerate areas.

MINOR STREAMS

Statement of problem.—In analyzing the general problem of correlation of minor streams with Triassic fanglomerates Mackin (1938) states:

> "The fact that the minor streams are closely spaced and that . . . conglomerates of various types are continuous for the greater part of the distance along the western margin of the Triassic Lowland makes the use of this criterion in the case of the small streams even more difficult than in the case of the major streams" (p. 39).

He further points out that where Triassic beds are in contact with the resistant crystallines of the Reading Prong minor streams flow southeastward to the basin, but where the Triassic is in contact with Paleozoic limestones the streams flow northward out of the basin (p. 39–40). In the region south of Bethlehem, Pennsylvania, and in the area between Reading, Pennsylvania, and the northeast end of the Carlisle Prong this obsequent type of north-flowing drainage is well developed. He concludes:

> "These relations, and the trellis drainage pattern that is well developed in parts of the Triassic Lowland, indicate that most of the minor streams of the area as a whole owe their positions and

directions of flow to headward erosion (in post-Triassic time) controlled by differential rock resistance." (p. 40)

Adequate discussion of possible relationships between Triassic fanglomerates and stream courses must include consideration of: (1) streams associated with fanglomerate deposits, (2) streams not associated with fanglomerates, and (3) fanglomerate deposits not associated with streams or with very minor tributaries. Due recognition must be given to the possibility that important outcrops may be obscured by glacial drift in the glaciated part of the Triassic basin. If glaciation has hidden critical evidence, proof of extraordinary correlation of minor streams with conglomerate areas manifestly cannot be obtained. Northeast of the Delaware transverse folding of the Triassic has shifted the border sediments laterally from their place of deposition. This factor alone prohibits positive correlation with present drainage.

Glaciated Triassic basin.—Immediately southwest of where the Hudson leaves the crystallines of the Highlands is the limestone breccia discussed in the section on the Hudson River. At present a small stream (1)[1] enters the basin near the southwest margin of the deposit. The asymmetric position of the stream in relation to the area of breccia suggests that the limited degree of matching is purely coincidental.

From 4 to 8 miles southwest of the above-described limestone breccia occurs the largest fanglomerate in the section. Near the fault line, midway between Ladentown and Wesley Chapel, the deposit consists almost entirely of angular limestone cobbles and boulders up to 4 feet in diameter (Pl. 2, Fig. 1). Basinward the fanglomerate contains more rounded pebbles of quartzite and some quartz in conglomerate ledges outcropping along the crests of ridges. No stream issues from the highlands opposite this fanglomerate.

The first minor river which enters the basin southwest of the Hudson is the Ramapo (2). There is no evidence of conglomerate directly opposite the mouth of this stream, but three-fourths of a mile northeast of the river, on the southern slopes of the diabase intrusion of Union Hill, is an outcrop of Triassic conglomerate which contains curvilinear to subrounded quartzite, sandstone, and quartz pebbles. Significant correlation of the Ramapo with Triassic conglomerate calls for a large area of conglomerate opposite the mouth of that stream. No such relation is observed.

Farther southwest the Wanaque and Pequannock rivers (3, 4) join to become the Pompton River near the place where the two enter the Triassic basin. Along the east side of Pompton Lakes about 1½ miles east of the Wanaque and Pequannock rivers is a small Triassic deposit of subrounded to subangular pebbles of limestone, sandstone, and quartzite. As in the case of the Ramapo, there is no area of conglomerate directly opposite the mouth of the valley, but rather a small deposit well to the east of the valley. Furthermore, the conglomerate is on the northeast limb of the Watchung syncline and dips 25° westward. Southwestward shifting of the conglomerate beds along the face of the highland border fault during post-Triassic folding could have moved the conglomerate from an original location even farther northeast of its present outcrop.

From 3 to 7 miles southwest of the town of Pompton Lakes is the second largest fanglomerate in the section, the Montville fanglomerate, which contains principally

[1]For ease of reference all the minor streams discussed in this chapter are numbered consecutively, from northeast to southwest (Pl. 1).

angular to subrounded gneissic and granitic pebbles and boulders in a coarse sandy matrix. The areal extent of this fanglomerate may be even greater than is shown on the map, since glacial debris probably covers part of the deposit. The only stream entering the basin from the highlands along the extent of the deposit is the very short creek (5) debouching at its extreme southwest end. The southwest trend of this stream and its small size indicate a post-Triassic subsequent origin.

The next large stream southwest of the Pequannock River is the Rockaway River (6). Wisconsin terminal moraine has blocked the preglacial valley of the Rockaway where the river once entered the Triassic basin. The river now enters the basin through a narrow gorge at Boonton, New Jersey, more than 3 miles northeast of its preglacial course.

Two small streams (7, 8) which enter the basin between the Rockaway River and the North Branch Raritan River are apparenly not matched by Triassic fanglomerates, although glacial debris may conceal conglomerate outcrops.

Unglaciated Triassic basin.—There is no fanglomerate in the Triassic where the North Branch Raritan River (9) enters the Triassic basin near the southwestern end of the Watchung trap ridges. The next minor river to the southwest is the Black River (10), the second largest stream in the section. Although the Black River does not flow across fanglomerate where it enters the Triassic basin, it is closely flanked on either side by deposits of quartzite and limestone breccia and conglomerate.

The North Branch Rockaway (11), the next stream to the southwest, enters the Triasssic basin near the northeast end of a deposit of limestone breccia. The extremely asymmetric position of this stream in relation to the breccia makes correlation of very doubtful significance.

Next to the southwest is the South Branch Raritan River (12), the largest stream in the section, which flows directly across a deposit of quartzite conglomerate as it enters the Triassic basin near Clinton. Most of the lower course of this river is along the limestone lowland of German Valley. This strongly suggests post-Triassic origin of the Raritan by subsequent erosion, in which case its matching with Triassic conglomerate would be purely by chance.

Southwest of the South Branch Raritan River is the largest Triassic border conglomerate in New Jersey, the Pattenburg, or Barrens, area of quartzite conglomerate. No stream enters the basin along the boundaries of the deposit. The eastern margin of the Pattenburg conglomerate lies 2½ miles west of the present course of the Raritan; its western margin is 8 miles from the Raritan. Correlation of this conglomerate with the nearest minor stream, the Raritan, involves postulating extensive shifting of the course of the Raritan. Actual field relations indicate that the Pattenburg deposit represents one of the most striking failures in the correlation of conglomerate areas with existing stream courses.

The three border conglomerates which flank the Delaware River where it flows into the Triassic basin are not associated with any minor streams entering the basin from the highland block.

The next area of conglomerate to the southwest has been correlated by Meyerhoff and Olmsted (1936, p. 38) with the Lehigh River and may therefore be omitted from discussion in this section.

A small stream (13) enters the Triassic basin east of Palm, near the southwest edge of a 7-mile-wide zone of quartzite-limestone conglomerate. The asymmetric position of this small stream in relation to the conglomerate does not favor generic relation.

The next three streams to the southwest entering the Triassic basin within this section (14, 15, 16) are not associated with fanglomerate. A breccia deposit lies between the second and third stream. The next stream to the southwest (17) matches an area of limestone breccia that appears on the Pennsylvania State geological map. The writer searched for this breccia deposit but could not find it. Next to the southwest is a very small area of breccia not matched by any stream. The next stream to the southwest (18) enters the Triassic area near the center of a limestone breccia deposit shown on the State geologic map, but which the writer could not find. The longest stream of the section (19), next to the west, flows for most of its length in a subsequent course along the east side of Oley Valley, a Paleozoic limestone lowland. It enters the Triassic near the center of a long strip of limestone breccia; correspondence with the breccia is good, but the subsequent course of the stream casts doubt upon its Triassic origin. The next stream to the west (20) flows along the west side of Oley Valley and enters the Triassic area at the extreme western end of the same strip of limestone breccia. The subsequent course of the stream and its asymmetric position in relation to the breccia deposit lend no support to the hypothesis of its Triassic origin. A very short distance west of the stream numbered 20 is a very small area of limestone breccia not matched by any stream. The next stream to the west (21) enters the basin where there is no breccia along the highland border. West of this stream is a limestone breccia deposit which extends for 3 miles along the border fault and terminates at the Schuylkill. No minor streams enter the Triassic area opposite this breccia deposit.

From the Schuylkill to the western end of South Mountain (the crystalline mass 8 miles west of Reading) no streams enter the Triassic basin from the north. Between South Mountain and the northeastern end of the Carlisle Prong lower Paleozoic limestone has been faulted into contact with the Newark series. The less resistant limestone forms a lower land surface than the Triassic, an obsequent fault-line scarp is developed along the line of contact, and nearly all of the minor streams now flow northward from the Newark area down to the Paleozoic land surface. The single exception is Swatara Creek (22), which flows from the Great Valley a short distance across the Triassic before joining the Susquehanna.

In the Triassic opposite Swatara Creek is a small deposit of conglomerate containing mainly curvilinear to subrounded quartzite and sandstone pebbles.

Three water gaps and one wind gap (lettered S, I, H, M, Pl. 1) which break the crest line of Blue Mountain, the Silurian quartzite ridge to the north, could hypothetically record the points at which streams of Triassic origin, now lost through capture, once crossed Blue Mountain before flowing along unknown courses into the Triassic basin where they could have deposited Triassic gravels.

Critical analysis of this type of correlation, however, indicates no results of value. Reconstruction of Triassic drainage lines through the four gaps can be based on an extremely wide range of possibilities, since each gap gives but a single point from which to swing each reconstructed stream course as widely as desired. The distance

from Blue Mountain is so great, and the potential courses for streams across the intervening distances are so numerous, that no confidence can be placed in any correlation attempted. Where there are so many possibilities a valid reconstruction of Triassic drainage is quite impossible, and correlations are equally impossible.

ANALYSIS OF CORRELATION

Within the glaciated section of the Triassic five streams (1, 2, 3, 4, 5) are only poorly matched by fanglomerates, two streams (7, 8) are not associated with fanglomerates, and the relations in the case of one stream (6) are completely obscured by glacial drift. The Hudson River is associated with only limestone breccia, sandstone, and shale. No stream is well matched by fanglomerates, although the blanket of glacial drift makes impossible valid conclusions on the precise degree of matching of streams with fanglomerates.

Outside the glaciated Triassic basin, between the North Branch Raritan River and the Susquehanna River, more valid conclusions may be drawn. Six minor streams (10, 12, 17, 18, 19, 22) and the Schuylkill show good matching with fanglomerate. Three minor streams (11, 13, 20) show only a very poor matching, and five minor streams (9, 14, 15, 16, 21) show no matching at all with fanglomerate. Both the Delaware and the Susquehanna are associated with relatively insignificant deposits of border breccia and conglomerate. Thus, 7 streams show good matching, and 10 show very poor matching or no matching with fanglomerate. Even allowing for the streams which show very poor matching, the degree of correlation is by no means extraordinary. Out of 17 streams, 12, or approximately 70 per cent, "hit", with widely varying degrees of accuracy, fanglomerate where they enter the Triassic basin. This is a better than even degree of correlation, but no better than it should be by the laws of chance. Between the North Branch Raritan River and the Susquehanna River the border fault spans some 128 miles. Fanglomerate lies opposite or fringes 84 miles, or about 66 per cent of the total distance.

It might be validly argued that only the larger of the Triassic minor streams and the major rivers would have persisted up to the present, and that the breccia areas should not be considered in this summary since they represent deposits of short-lived Triassic streams gullying newly uplifted facets of the border fault. If only the relations of conglomerate deposits to streams equal to or larger than the size of the North Branch Raritan River are analyzed, different results are obtained. Three minor rivers (10, 12, 22) and the Schuylkill show good matching with conglomerate, and two minor rivers (9, 19) and the Delaware and Susquehanna rivers show none at all. Four streams "hit" conglomerate where they enter the basin; four do not—a 50 per cent degree of matching. This lower degree of matching is very close to what it should be by chance, since, out of a total distance of 128 miles of border fault, conglomerate lies opposite 61 miles, or 48 per cent of the total distance.

Correlation of all the streams with conglomerate and breccia, or major and minor rivers with conglomerate, is not only far from extraordinary but is no better than is to be expected through purely chance association of many closely spaced streams crossing a fault line which is bordered by extensive stretches of fanglomerate.

In field conference and in correspondence, Meyerhoff stated that the extraordinary correlation between modern streams and Triassic fanglomerates which Meyerhoff and Olmsted believe to exist was secured in part, at least, by postulating former stream courses different from those now present. Salient examples of these follow.

Although the Lehigh does not now enter the basin a course for the river was postulated which would connect it with an area of conglomerate. In field conference Meyerhoff suggested that during the Triassic the Schuylkill entered the basin 4 miles west of its present position. The Susquehanna was correlated with pebble beds now outcropping more than a mile west of the river.

Meyerhoff has postulated that the North Branch Raritan River did not enter the basin during the Triassic at its present point of entry (where no conglomerate outcrops) but that it originally flowed southwest parallel to the present boundary fault, depositing the Mt. Paul conglomerate and the next large area of conglomerate more than a mile southwest of Mt. Paul.

Southwest of where the South Branch Raritan River enters the basin is the largest Triassic border conglomerate area in New Jersey, the Pattenburg, or Barrens, area of quartzite conglomerate. No stream enters the basin along the boundaries of the deposit. Meyerhoff has postulated that the Pattenburg conglomerate was deposited by the South Branch Raritan, although the nearest position of the conglomerate lies 2½ miles west of the present course of the Raritan.

Meyerhoff has also suggested that the Furnace Hills conglomerate might be correlated with wind and water gaps in Blue Mountain—thus involving postulation of stream courses across the Great Valley area where none now exist.

The case of the Hudson River involved a slightly different method of correlation. The complete lack of any apparent genetic relationship between the Hudson and the Triassic sediments with which it is now associated has been explained by postulating that the Hudson did not come into existence until after Triassic time.

Meyerhoff's postulate (personal communication) of former stream courses different from those occupied now is based upon the supposition that fanglomerates deposited by streams in Triassic time would have proved so resistant that the streams would have since taken courses marginal to the conglomerate masses.

Only one stream course in the Triassic basin northeast of the Carlisle Prong could fit this hypothesis—the Schuylkill. The Schuylkill enters the basin in a south-flowing course and then turns abruptly to the southeast in a course marginal to the Robeson conglomerate. This river has been listed by the writer as an example of one of the "hits". The other streams which fail to "hit" conglomerate have courses which give no suggestion of turning at their entrances into the basin and taking courses marginal to conglomerate masses. If, on the other hand, the whole course of each of these streams has changed since Triassic time, no proof remains that they ever flowed in the courses postulated by Meyerhoff.

It is a fundamental principle of correlation studies that, when a correspondence in time or space between two sets of phenomena is used to prove a genetic relationship, the degree of precision of the correspondence determines its probative value. A low degree of correspondence has no significance, since it may be purely the result of

chance. A high degree of precision, far too high to be the product of chance, alone has probative value. Meyerhoff and Olmsted appear to have not realized that, in postulating for a number of streams Triassic courses widely different from those actually observed, in order to bring these streams into the Triassic basin closer to local exposures of conglomerate, they were admitting repeated failures in correlation and destroying the probative value of any correlation obtained by such procedure.

SUPPOSED TRIASSIC VALLEYS AND RE-ENTRANTS IN THE HIGHLAND BORDER

Triassic-filled, stream-cut re-entrants which notch the fault block bounding the Triassic basin on the northwest constitute a second line of evidence in favor of a Triassic and pre-Triassic age of the present drainage. Meyerhoff and Olmsted describe these re-entrants as follows:

> "Most of these streams—possibly all, but the heavy glacial veneer makes a sweeping statement unsafe—enter the Triassic lowland through V-shaped notches and re-entrants cut in the fault-line scarps. The notches are normal fluvial forms, but the presence of Newark sediments in those wherein outcrops project through the glacial drift and post-glacial wash, shows that they are Triassic, and not Cenozoic, in origin" (1936, p. 38–39).

They (1936, p. 39) recognize that such re-entrants might have been reoccupied by post-Cretaceous streams as a result of differential erosion. Thus their presence would not necessarily prove Permo-Triassic origin of present drainage. Proof that the original streams still occupy the re-entrants is based on a supposed extraordinary correlation between the streams and Triassic conglomerates. Since this correlation is no stronger than a normal chance occurrence, the re-entrants, if present, do not prove Permo-Triassic origin of the drainage.

The localities of the re-entrants discussed below are seven examples presented by Meyerhoff and Olmsted during a field conference. Of these seven re-entrants, two were based on reported outcrops that could not be relocated during field examinations with Doctors Meyerhoff and Olmsted. These were a re-entrant in the moraine-filled mouth of the preglacial valley of the Rockaway River and a re-entrant west of the Schuylkill between Reading and Shillington.

A third re-entrant was located where the Ramapo River emerges from the Highlands. Here the only topographic evidence for the possible presence of a stream-cut re-entrant is a slight offset in the Ramapo Mountain front. Similar offsets at many other points can be shown to be due to en-echelon faulting. Taken by itself it does not prove an origin through Triassic stream erosion. Within the offset, however, an outcrop of Triassic sandstone and of trap had been reported. Neither could be located on a joint field trip with Meyerhoff and Olmsted. An outcrop that may have been responsible for the reported existence of trap turned out to be smoke-blackened gneiss.

The fourth re-entrant, where the Wanaque and Pequannock rivers enter the Triassic basin, was apparently based on an irregularity in the Ramapo Mountain front caused by infaulting of Hudson shale between the Triassic sediments and the crystalline mountain front.

From about 6 to 18 miles southwest of Boonton is a secondary fault block in front of, or southeast of, the main highland block. The secondary fault block, here called

the Morristown block, is about 12 miles long in a northeast-southwest direction and about 5 miles wide. Its southeastern margin forms part of the boundary of the Triassic basin. That the main boundary fault of the highland block forms the northwest boundary of the Morristown block is shown at the southwest end of the Morristown block, where a wedge of Hardyston sandstone, Kittatinny limestone, and Martinsburg shale has been infaulted between the pre-Cambrian segments of the two fault blocks. Northeast of the wedge of Paleozoics the fault-line scarp continues as a clearly marked topographic feature, although it is bounded on both sides by pre-Cambrian crystallines. On the Martinsburg shale of the Paleozoic wedge is a high hill, Mt. Paul, which is capped by Triassic quartzite conglomerate. The conglomerate at Mt. Paul, interpreted by Meyerhoff and Olmsted as resting in a downfaulted subsequent valley, is one of the seven Triassic re-entrants. Infaulting of a strip of Paleozoic and Triassic sediments along well-known and long-recognized faults which cut both Paleozoics and Triassic, followed by post-Triassic erosion, accounts for all the observed facts.

Overlap of Triassic on Paleozoic shale and limestone between Perryville and Clinton, New Jersey, at the South Branch Raritan River is the evidence for another re-entrant. This re-entrant is located where a great fault brings Stockton and Lockatong beds to the surface adjacent to the highland boundary fault. The base of the Triassic is exposed, as well as the underlying Martinsburg shale and Kittatinny limestone on which the Triassic rests. Finally, the basement complex upon which the Paleozoics rest is exposed opposite the horns of Cushetunk Mountain. The mere fact of overlap constitutes no evidence for a re-entrant, since overlap covers interstream areas as well as valleys and may cover surfaces never dissected by stream erosion. Association of the South Branch Raritan with the overlap does not constitute a V-shaped re-entrant cut in the fault-line scarp.

The last supposed re-entrant, at the Susquehanna River, was inferred from the northward arching of the Paleozoic-Triassic contact between the Reading Prong and Blue Ridge, and on overlap of Triassic on Paleozoics an unspecified distance west of the river. The northward arching of the contact is due to irregularity in the trend of the border fault. As in the case of the South Branch Raritan River overlap does not prove a Triassic valley. Since the overlap is west of the river it has no bearing on the problem of whether or not the Susquehanna enters the Triassic lowland through a V-shaped notch or re-entrant cut in the fault-line scarp.

It is concluded that field evidence does not show that Triassic sediments occupied any of the present stream valleys cut in the Highland crystallines or other older rocks north and northwest of the Triassic basin.

STRUCTURAL AND STRATIGRAPHIC CONTROL OF GRAVEL LITHOLOGY OF THE TRIASSIC BORDER FANGLOMERATES

GENERAL STATEMENT

The primary structures of the Triassic border conglomerates indicate deposition by short streams of high declivity. There is no evidence in the degree of rounding of the gravels of the conglomerates that some of the streams which deposited them were

long, master streams, and some shorter, minor streams. Rather, the conglomerates all contain gravels showing about the same degree of rounding, and they indicate deposition by streams of about the same length. Their observed degree of rounding could be attained in less than 10 miles of transportation. Since the physical character of the conglomerates points to deposition by short, steep streams, it remains to be determined whether the variations in lithologic character of the border fanglomerates—conglomerates and breccias—can be accounted for by a hypothesis of local derivation.

CONTROL BY STRUCTURES OF THE NORTHWEST HIGHLAND BLOCK

Between the Triassic basin and Blue Mountain rise the major structural uplifts of the Carlisle Prong, or South Mountain, and the Reading Prong. The major folds involved in the Carlisle Prong uplift affect formations as young as the Ordovician Chambersburg limestone; the Chambersburg and the underlying Beekmantown limestone form a long anticlinal nose trending north of east and extending beyond the Susquehanna River as far east as Swatara Creek. The Reading Prong uplift is a faulted anticlinal mass exposing wide areas of pre-Cambrian rock, whose continuous outcrop is interrupted only by transverse sags at Bethlehem and Reading, Pennsylvania. The southwestern end of the Reading Prong is at South Mountain, 12 miles west of Reading. West of South Mountain the uplift, now a simple anticline faulted at its southern margin by the Triassic border fault, continues past the Susquehanna River. Thus, in the area of the Susquehanna the ends of the Carlisle Prong and Reading Prong uplifts overlap and parallel each other, separated by a narrow syncline preserving Martinsburg shale between uplifted bands of Beekmantown and Chambersburg limestone. So close together are the ends of the Carlisle Prong and Reading Prong uplifts that the overall structural effect is of one major anticlinal uplift with a wide transverse sag in the Susquehanna River region and minor sags at Reading and Bethlehem.

If movement along the Triassic border fault could drop the Triassic basin several hundred feet at the present time, debris eroded from the resulting escarpment would vary widely in lithologic character from point to point along the fault. There would be a systematic arrangement of rock types deposited in the basin as controlled by the sags and upwarps in the Carlisle Prong-Reading Prong uplift, since these transverse structures in the main uplift affect the outcrop of rocks of pre-Cambrian to Ordovician age along the Triassic border fault. It follows that this same control should have affected the lithologies of gravels deposited in the Triassic basin during Triassic deposition. Obviously, the control could be found only in those parts of Pennsylvania where Newark structures are largely homoclinal and where Triassic beds an equal distance from the border fault are approximately the same age.

Near the northeast end of the Carlisle Prong in the Fairfield and Gettysburg region, Stose and Bascom (1929) have mapped the Arendtsville fanglomerate lentil, a border fanglomerate which stretches for 20 miles parallel to the border fault. The fanglomerate contains fragments of Cambrian quartzite and sandstone and pre-Cambrian aporhyolite. Farther northeast the border fanglomerates are made up of Cambrian and Ordovician limestone fragments; these limestone breccias extend parallel to the border fault as far northeast as Dillsburg. Still farther east, between Lisburn and

a point a short distance east of Swatara Creek, border fanglomerates contain fragments derived from Silurian and Devonian rocks. Thus, northeastward and eastward down the transverse sag in the Carlisle Prong-Reading Prong axis Triassic border fanglomerates are derived from a progressively younger group of formations—from pre-Cambrian to Devonian.

The center of this sag is somewhere near where Swatara Creek enters the basin. To the east is a great area of fanglomerate derived from the Devonian Oriskany or Esopus, Silurian quartzite and sandstone, and Cambro-Ordovician limestone. Still farther east where the Reading Prong uplift rises sharply at Reading, only Cambro-Ordovician limestone is represented in border fanglomerates. The sequence along the west side of the sag has in large part been repeated in reverse order. In the Allentown-Bethlehem region the outcrop of the Reading Prong crystallines markedly narrows, denoting a secondary sag in the anticlinal axis. Opposite this sag Silurian quartzite and sandstone gravels reappear in the border fanglomerates. The correlation between variations in uplift of the Carlisle Prong-Reading Prong anticline and the lithologies in the Triassic border fanglomerates is more systematic than is readily explained by chance.

This correlation is more explicitly shown by the cross section in Plate 1, on which has been constructed the erosion surface cut in the northwest highland block in late Newark time during deposition of Newark sediments now preserved about 1 to 2 miles out in the basin. The line followed by the section runs along the axis of the long uplift from the northeast end of the Carlisle Prong to a short distance northeast of Bethlehem, Pennsylvania. The section ends at this point since uplift by the Flemington fault of New Jersey exposes next to the border fault Brunswick beds of progressively older age from near the Delaware River toward the northeast, and therefore Triassic beds an equal distance from the border fault are no longer of the same age.

The section has been constructed from sections published by Miller (1939b, p. 7) for the Lehigh Valley region; Willard (1939, p. 5, 6), and Willard and Fraser (1939, p. 19) for the Reading district and the Schuylkill Valley region; Willard and Cleaves (1938, Pl. 10), Willard, Swartz, and Cleaves (1939, Figs. 40, 93) for the Susquehanna Valley region; and Stose and Bascom (1929) for the South Mountain region.

A basic and necessary assumption is that stratigraphic thickness and lithologic character are essentially constant from the points where sections are used to the line of reconstructed section. The average thickness shown for the Martinsburg shale is about 3000 feet (Stose, 1930, p. 634). This figure was chosen because all sections described in the area between the Susquehanna River and the Schuylkill give 3000 to 3500 feet as the thickness of the Martinsburg. Behre (1933, p. 137) has estimated the thickness of the Martinsburg as about 11,000 feet in the Lehigh Valley region.

Stose and Jonas (1935a; 1935b; 1939a) and Stose and Stose (1940) claim that the Reading Hills are part of a great overthrust sheet whose roots are under the Triassic basin. Such a structure is not shown in Plate 1, inasmuch as Miller and Fraser (1935), Fraser (1938), and Miller (1944) maintain that the presence of an overthrust here has not been proven.

Raymond (1927, p. 246–249; 1939, p. 147) holds that Newark climate was warm

and moist. It is therefore postulated that quartzite and sandstone in the highland block would have formed a mountainous terrain, while limestone and shale would have formed lowlands.

The cross section is necessarily idealized since the stratigraphic data involved are derived from sections made in localities up to 18 miles from the line of cross section. Most of the thicknesses shown are therefore original estimated thicknesses transposed across wide areas where the units may be significantly thicker or thinner. Mechanical obstacles to showing smaller structures on such a scale further generalize the diagram. The Pennsylvania State geological map and more detailed maps of the highland block interpret the Reading Prong as a highly faulted anticlinal mass, but it is here drawn with smooth contours.

Two minor unconformities, one at the base of the Jacksonburg and one at the base of the Martinsburg, and the profound Taconic unconformity at the base of the Silurian further complicate the problem of a valid reconstruction of the Paleozoics overlying the highland block pre-Cambrian. The thickness of the Cambrian and Ordovician section which was removed by pre-Tuscarora erosion is unknown.

In short, the reconstruction cannot explain local and minor variations in lithology of the gravels but is advanced to explain the larger variations of gravel lithology.

Comparison of the reconstructed erosion surface with Triassic border fanglomerates discloses that near the northeast end of the Carlisle Prong the Triassic erosion surface would have truncated Cambro-Ordovician limestone opposite that part of the border fault where Triassic fanglomerates contain only Cambro-Ordovician limestone fragments. The Silurian Tuscarora and Clinton quartzite and sandstone and the Devonian Hamilton (Montebello) sandstone would have formed high ridges in he general area of the Lisburn conglomerate, which contains gravels derived from Silurian and Devonian formations. Devonian limestone occurs opposite that section of the fault line where Triassic limestone breccia deposits now border the fault. The Hamilton sandstone would have outcropped on the eastern limb of the sag in the general area of the conglomerate which outcrops where Swatara Creek enters the basin.

The section east from Swatara Creek to Furnace Ridge, where Triassic border sediments are predominantly shale with interbedded sandstone, would lie opposite the shaly Bloomsburg red beds and the lower part of the mountainous terrain formed by Silurian sandstone and quartzites. A great anticline and syncline of Clinton and Tuscarora sandstone and quartzite would form mountainous terrain directly opposite the great Robeson conglomerate area and its western extension, the Furnace Ridge conglomerate, in which Silurian sandstone and quartzite cobbles are the main constituents. Only at this one place along the highland border block would the combination of an anticline and syncline permit such high mountains to be formed on Silurian quartzite and sandstone directly opposite the largest area of Triassic quartzite and sandstone conglomerate in the basin. The upwarp of the main part of the Reading Prong east of Reading would expose Cambro-Ordovician limestone in lowlands opposite Triassic border sediments which consist of Brunswick shale with limestone breccias fringing the border fault. The transverse sag in the Prong in the Allentown-Bethlehem region would have permitted preservation of Silurian Tuscarora and Clinton beds in the highland block opposite the same general area

where Triassic border fanglomerates of Silurian sandstone and quartzite gravels now outcrop.

LOCAL WEAKNESSES OF THE HYPOTHESIS OF STRUCTURAL CONTROL

Oley Valley, a downfaulted re-entrant in the face of the Reading Prong from 6 to 10 miles east of Reading, Pennsylvania, is not shown on the cross section. In Oley Valley Martinsburg shale and Cambro-Ordovician limestones are downfaulted against pre-Cambrian rocks. If this downfaulted block, which directly borders on the Triassic basin, was in existence during Newark sedimentation, the Newark border sediments adjacent to the block should contain fragments of greatly different stratigraphic origin from Newark sediments east and west of the block. No such relations are observable. Breccia deposits derived from Cambro-Ordovician limestone fringe the border fault from Palm southwestward to Oley Valley. One such deposit extends from 2 miles east of Oley Valley across the mouth of the valley, with no lithologic change. This suggests that the downfaulted block of Oley Valley was not in existence during Triassic deposition but originated during the post-Newark Palisade disturbance, when the many normal faults of the Reading Prong probably formed (Bayley, Salisbury, and Kümmel, 1914, p. 19, 21). The fault block is therefore omitted from the cross section. If, however, the fault along the west side of Oley Valley predates the Triassic sediments opposite the valley, the hypothesis of control of fanglomerate lithology by structures of the Reading Prong axis locally fails to account for the lithologic character of the fanglomerate at Oley Valley.

A pebble of Devonian Esopus or Oriskany sandstone was found in the upper part of the Robeson conglomerate. Inasmuch as about 1875 feet of upper Silurian and basal Devonian beds separates the Clinton sandstone from the Oriskany, the reconstruction of Paleozoics mantling the highland block cannot account for Oriskany beds in the syncline at Reading. Devonian beds could have outcropped no closer than the present area of outcrop of the Martinsburg shale.

Lawrence Whitcomb (personal communication) reports an arkosic matrix in fanglomerate at Hosensack, Pennsylvania, about 3 miles northeast of Palm. The writer found a granite fragment in the Triassic near Grill, Pennsylvania, about 2 miles south of Reading. The reconstruction fails to account for these occurrences of crystalline material.

The lithology of the Oley Valley fanglomerate can be explained by a post-Triassic fault, the Devonian pebble in the Robeson conglomerate by an entirely expectable moderate variation in length of streams eroding the highland block, and the crystalline material in the Robeson conglomerate and the fanglomerate at Hosensack by deep-cutting streams reaching exceptionally high portions of the basement complex very late in Triassic sedimentation. Although these problems, which require special explanation, reduce the degree of precision of correspondence between fanglomerate lithology and highland block structures, the following considerations indicate that the hypothesis is not disproved.

(1) The idealized nature of the cross section and the uncertainties of true thickness of formations shown limits the section to explanation of only the larger variations in gravel lithology, and thus cannot explain local and minor variations.

(2) The good correspondence between the larger variations in gravel lithology and

structures of the highland block already described suggests that the principle involved is essentially valid.

EFFECT OF PROGRESSIVE STRIPPING OF PALEOZOICS FROM THE NORTHWEST HIGHLAND BLOCK

If there were no transverse folding of the Triassic northeast of the Allentown-Bethlehem region, the continued widening of the Prong which suggests rising of the basement floor northeast from Bethlehem should result in the appearance of Triassic gravels derived from progressively older Paleozoics toward the northeast. Ultimately, only pre-Cambrian-derived gravels would occur in the fanglomerates.

Complications arise, however, that prevent this orderly sequence: (1) Whereas the Prong widens, infolding of Paleozoics during the Appalachian revolution locally lowered the structural height of the Prong. (2) Northeast of the Rockaway River in the Green Pond Mountain region Silurian quartzite overlaps some 6000 feet of Cambrian and Ordovician sediments to rest directly upon the basement complex. (3) A new quartzite, the Devonian Skunnemunk conglomerate, which is quite similar to the Silurian conglomerate, appears in the Green Pond Mountain region, about 2200 feet stratigraphically above the Silurian quartzite. (4) The Triassic itself is transversely faulted and folded to expose beds of widely varying age next to the border fault.

For example, in the Pattenburg-Clinton area the Flemington fault has raised to the surface the basement of the Triassic, the Stockton, Lockatong, and lower and middle Brunswick beds directly opposite the highland block. Border fanglomerates of Stockton, Lockatong, and lower Brunswick age here are composed principally of Silurian, or possibly Devonian, quartzite. Inasmuch as the nearest outcrop of the Skunnemunk quartzite is more than 40 miles to the northeast and there is no evidence that it ever extended as far southwest as this region, it is suggested that the source beds for the conglomerates are Silurian rather than Devonian. If so, then throughout Stockton, Lockatong, and early Brunswick time Silurian quartzite held up high hills in this part of the Prong which were for a considerable time a source of quartzite gravels. The Prong in this area is a major anticline with minor downfolded and faulted segments. The anticline pitches southwestward, and the Silurian quartzite would thus have pitched southwestward in this region. Erosion of the pitching structures would have caused progressive displacement of the outcrop down the pitch, resulting in deposition of quartzite gravels in the Triassic border sediments at progressively higher horizons toward the southwest. This effect can be seen in the Triassic border fanglomerates southwest of Clinton, New Jersey. Elongation of the Triassic Pattenburg fanglomerate toward the southwest and the presence of quartzite conglomerates on both sides of the Delaware River may be due to this process.

The Watchung Mountains outline the border of a broad, shallow, transverse fold. A secondary but smaller syncline occurs at New Germantown and is marked by the curving outlines of the New Germantown basalt sheet. Between these synclines is an anticline in the Triassic. The combination of two transverse synclines separated by an anticline results in exposure at the highland border of Triassic sediments of varying Brunswick age. The effect of progressive stripping of Paleozoics from the

Prong region should result in the presence of younger Paleozoic fragments in the anticlines where older Triassic beds are exposed, and older Paleozoic materials in the synclines. Opposite the New Germantown syncline occurs a Triassic breccia deposit of Cambro-Ordovician Kittatinny limestone fragments, while in the anticline to the northwest is exposed quartzite fanglomerate chiefly derived from Silurian quartzite or possibly Devonian Skunnemunk quartzite.

At the northeast end of the Watchung fold, near Darlington, New Jersey, a boulder of Helderberg limestone has been found. Near Pompton Lakes, higher in the Brunswick sequence, occurs a conglomerate consisting of Kittatinny limestone; dark-gray sandstone, possibly Devonian Kanouse; and red quartzite conglomerate of either Silurian Green Pond or Devonian Skunnemunk origin. Near the center of the broad syncline at Montville, in the highest Brunswick beds, occurs a Triassic conglomerate deposit composed principally of pre-Cambrian crystalline cobbles and boulders. The crystalline conglomerate owes its present lithologic character and location to its stratigraphically high position preserved in or near the center of the Watchung Mountains syncline and to the structural height of the Reading Prong uplift, which was sufficiently high for Triassic streams cutting in the highland block to reach the basement complex late in Triassic sedimentation. The two conglomerate outcrops to the northeast of the Montville area illustrate progressive stripping of the Paleozoics from the Prong previous to late Brunswick time when the basement complex was uncovered. Farther southwest, in Pennsylvania, the massive ledges of quartzite conglomerate in the upper part of the Robeson conglomerate appear to represent the most active period of denudation of the Silurian sandstone-capped anticline in the highland block to the north, while the uppermost Robeson conglomerate beds, which are high in limestone pebble content, and the limestone breccia deposits north of the Robeson conglomerate would represent the deposits formed after the anticline had been unroofed to expose Cambro-Ordovician beds.

SUMMARY

The present distribution and lithologic character of Triassic border fanglomerates southwest of the Delaware River can be largely accounted for by a reconstruction of a Paleozoic terrain on the northwest highland block during late Newark sedimentation.

Northeast of the Delaware transverse faulting and folding of the Newark series place Newark fanglomerate of widely varying age adjacent to the northwest highland block, revealing evidence of progressive stripping of Paleozoics from the highland block throughout Newark sedimentation. The effect of progressive stripping of Paleozoics from the highland block can also be noted in the Triassic border sediments southwest of the Delaware River.

EVIDENCE OF NEWARK LITHOLOGY

Meyerhoff and Olmsted have postulated an eastern or southeastern regional drainage and, necessarily, a southeastern regional slope in the Appalachians by the beginning of Newark time. They state:

"... modern drainage on the Atlantic slope has evolved, directly ... from the initial fluvial systems which drained the Appalachian Mountains *immediately* [italics by present writer] after they were folded in Permian time. (1936, p. 24) .... Trunk streams established east-flowing courses by utilizing structural sags and transverse faults *before the Newark epoch*" [Italics by present writer] (1936, p. 21).

If the main trunk streams of the Appalachians were established in southeastern courses by Newark time and if the regional slope of the Appalachians was southeastward at the beginning of the Newark, this drainage system should have strongly influenced sedimentation in the newly formed Newark basin. Apparently, Meyerhoff and Olmsted recognize this consequence of their hypothesis, for they state:

"Stose has claimed that only late Newark sediments display a Paleozoic-Appalachian parentage and that early and middle Newark formations reveal the derivation of their clastic constituents from the southeastern Piedmont. As a general statement Stose's claim cannot be sustained. The Stockton formation (lower Newark) is usually exposed near the Piedmont margin where the compositional influence of the local crystalline rocks is both plain and expectable. Lower and middle Newark strata reach the surface at few localities situated near the northwestern fault-line scarp, but in these localities clastic constituents of Paleozoic derivation dominate, and in the upper Newark deposits the Paleozoic influence is strong" (1936, p. 37).

Meyerhoff and Olmsted obviously are referring to those Stockton outcrops along the upthrown side of the Flemington fault, which are the only occurrences of Stockton near the northwest margin of the basin.

The Stockton is arkosic in its exposures along the Flemington fault where it crosses the Delaware River within 10 miles of the northwest border of the Newark, or more than half way across the Newark basin. The Stockton beds can be traced even closer to the northwest border of the basin along the same fault. South of Clinton, New Jersey, the Stockton arkose extends to within 7 miles of the Reading Prong. The distance to the border is even less before Paleozoics dominate the Stockton lithology. (Bayley, Salisbury, and Kümmel, 1914, p. 14) Previous workers on the Triassic of the Newark basin have held that the arkose of the basal Triassic was derived from a crystalline highland area southeast of the basin.

Thus, in the one place in the basin where the relative influences of the northwest border block and the southeast highland on the sedimentation of the Stockton can be evaluated, the arkosic phase of the Stockton persists for more than three-quarters of the way across the basin. Farther west, at the Delaware River, the arkosic phase of the Stockton continues well over half way across the basin; that distance is a minimum.

If the regional slope of the Appalachians was southeastward at the beginning of Newark sedimentation, the relative importance of northwest highland-block sedimentation as compared with southeast highland sedimentation should be just opposite to that observed. If the Delaware River was in existence at the beginning of Newark sedimentation and was heavily loaded with the erosional debris of the high Appalachians to the northwest, its contribution to the alluviation of the Newark basin would be expected to extend more than 10 miles out into the basin during the deposition of the 2300 to 3100 feet of Stockton beds.

Instead, the trans-basin lithologic character of the Stockton of New Jersey provides the clearest and most striking evidence for higher lands southeast of the basin at the beginning of Newark sedimentation. The middle and upper Newark beds also attest

to the same paleogeographic conditions. Stose and Stose (1944, p. 77) have stated that the quartz, feldspathic sand, clay, mica flakes, and red iron oxide which form the sands and clays of the Newark group were derived from the southeastern highlands. According to Kümmel (Lewis and Kümmel, 1940, p. 107), "... the great bulk of the feldspathic and micaceous sandstones that make up so much of the Newark rocks must have come from higher lands that still existed to the southeast."

These relations suggest that, until the time of depression of the Newark belt of sedimentation, drainage across the Appalachians was northwestward. Sinking of the Newark belt and rising of its northwest highland block disrupted the courses of these northwest-flowing streams and diverted them into courses longitudinal to the basin. The newly formed consequent streams which drained the northwest highland block made up the first southeast-flowing drainage system in this section of the Appalachians.

CONCLUSION

No extraordinary correlation was found between the highland border sediments of the Newark series and the present drainage of the central and northern Appalachians. There is, further, no evidence of major stream deposition of the highland border fanglomerates, nor evidence that any present-day stream enters the Newark basin through Triassic-filled remnants of Triassic valleys. Thus, the highland border sediments of the Newark basin fail to provide evidence for Triassic or pre-Triassic origin of the present drainage.

On the contrary, a number of lines of evidence indicate that at the time of Triassic deposition the streams which entered the Newark basin from the northwest were relatively short streams which originated as consequents upon the steep flexure or fault which first formed the northwest margin of the subsiding Newark basin. The Triassic fanglomerates bear all the characteristics of deposition by short streams of high declivity. The degree of rounding of the gravels in the fanglomerates could have been caused by only 5 miles of mountain stream transportation. The source beds of definitely identified Paleozoic materials in the Triassic gravels even now lie within 20 miles of the Triassic basin and in Triassic time must have outcropped even closer to the basin. The good correlation between lithologic character of the gravels and structures of the northwest highland block and between stratigraphic position of the gravels in the Triassic and the age of their source beds further supports this hypothesis.

The hypothesis accords with the lithologic character of the Newark group which indicates dominant sedimentation in the Newark basin by streams heading in highlands still existing to the southeast.

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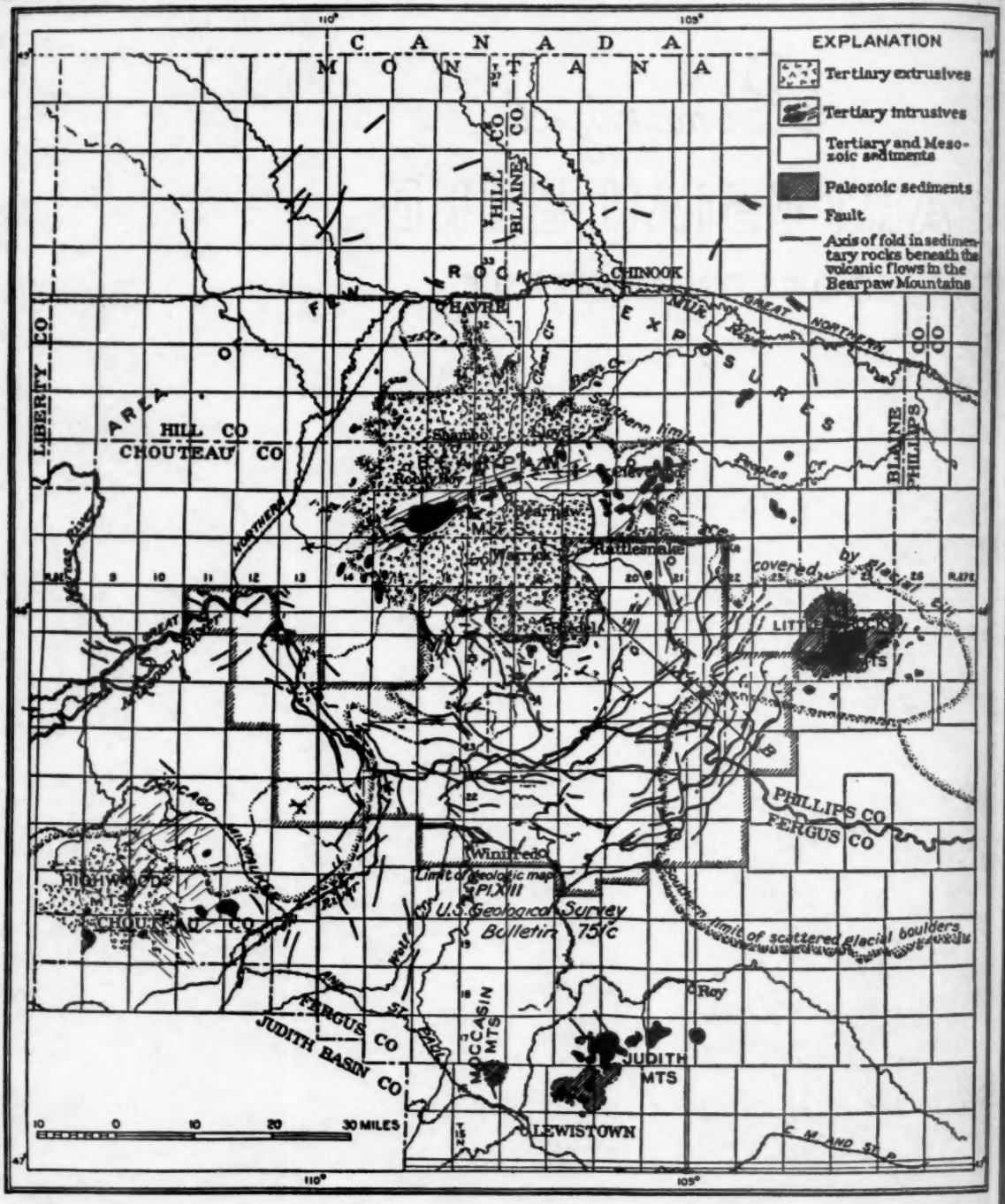

MAP SHOWING THE EXTENT AND TREND OF THE FAULTS ADJACENT TO THE BEARPAW MOUNTAINS

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ORIGIN AND MECHANICS OF THE THRUST FAULTS ADJACENT TO THE BEARPAW MOUNTAINS, MONTANA.

BY FRANK REEVES

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ABSTRACT

The hypothesis is presented that the thrust faults in the plains adjacent to the Bearpaw Mountains in north-central Montana were produced by a plainsward sliding of the volcanic rocks and underlying strata down the flanks of the low sedimentary arch on which the volcanics rest. It is predicated that this slipping took

place on a bed of bentonite in the Colorado shale and was aided by earth tremors that accompanied the last stages of the volcanic outbursts. The flat-lying Upper Cretaceous and lower Tertiary strata in the adjacent plains which, if more rigid, would have obstructed the plainsward sliding of the tilted mountain mass or been overridden by it were shoved plainsward and thrust-faulted for a distance of 20 to 30 miles.

The faults show no progressive plainsward decrease in amount of thrust across the broad area in which they occur and are separated by wide belts of undeformed flat-lying strata. These anomalous features are attributable to slight frictional resistance to movement offered by a slippery bentonite bed and the probability that the sedimentary rocks over part of the faulted belt were overlain and strengthened by a cover of volcanic rocks that thinned toward the outer margin of the faulted belt. The deformed strata consist chiefly of shales that were too weak to support the load involved in lifting the strata in anticlinal folds. Therefore, thrust faulting, instead of anticlinal folding, took place.

INTRODUCTION

The thrust faults in the plains south of the Bearpaw Mountains have been described by the writer in previous publications (1924a; 1929; 1924b; 1925). The two latter publications present the evidence tending to show that the thrust faulting is confined to Upper Cretaceous and Lower Tertiary formations. The information obtained during the past 25 years in drilling for oil and gas adjacent to the thrust faults supports the concept of shallow faulting, so that this feature of the structure seems to be fairly well proved.

Pre-Cretaceous strata were formerly assumed to be included in the thrust movement, reacting so as to allow crustal shortening without the formation of folds or large-scale thrust faults. This assumption was based on the fact that 80 per cent of the faulted strata are soft shales and were therefore thought to be too weak to transmit the thrust force for any considerable distance. Moreover, the uniformity in the throw of the faults and their separation by wide belts of undisturbed strata appeared inexplicable on the hypothesis that the thrust force was transmitted by the surface strata, whereas such features presumably could be produced by the drag effects accompanying a shortening of the underlying rocks. The assumption of a low coefficient of friction which would allow a shallow part of the crust to slide readily over its base and produce the unusual structural features characteristic of the faulted area however was not then considered.

Four hypotheses for the origin of the thrust faulting were discussed by the writer (1924b)—, but none was definitely adopted. After further study, one of the earlier hypotheses—that of a plainsward sliding of the mountain mass—has been accepted as the cause of the faulting. This article presents a mechanical analysis of the proposed hypothesis and a discussion of the extent to which it explains the unusual features of the thrust faults and related phenomena. A brief review of the main structural features in the faulted belt and the Bearpaw Mountain volcanic center is first presented. The position of the Bearpaw Mountains in relation to the Rocky Mountain front is shown in Fig. 1.

REVIEW OF THE STRUCTURE OF THE THRUST-FAULTED AREAS

The thrust faults on the south side of the Bearpaw Mountains occupy an arcuate belt 20 to 30 miles wide (Pl. 1). A belt of similar thrust faults occurs in the plains to the north of the mountains, but because of glacial cover the faults are poorly

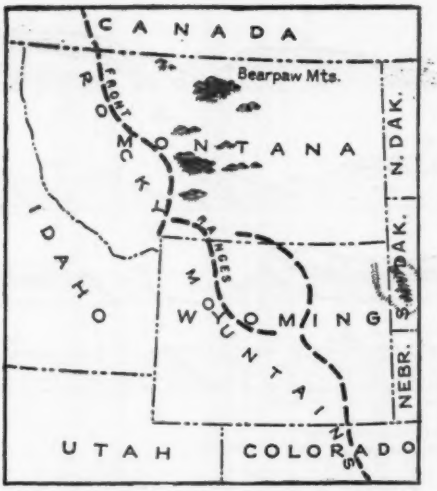

FIGURE 1.—*Index map showing location of the Bearpaw Mountains, Montana*

exposed. Few, if any, thrust faults are present in the plains immediately to the east and west of the mountains. Similar thrust faults in the plains on the northeast side of the Highwood Mountains are thought to be due to a plainsward sliding of the Highwood Mountains.

The faults are best exposed south of the Bearpaw Mountains, where they have been mapped in detail by the writer (1924a). Many of the faults are 10 to 20 miles long and are circumferential to the outer margins of the mountains. A few radial faults intersect the circumferential faults at right angles and appear to offset them. There are also several faults that do not conform to either of the above trends. The stratigraphic throw ranges from 700 to 1600 feet, and many faults attain the maximum throw in the outer margin of the faulted area.

The thrust faults are marked by a zone of upthrust strata 1000 to 3500 feet wide. The variation in width of the upthrust strata is due to a slight regional tilt to the eastward so that the thrust faults are eroded to progressively lower levels from east to west across the faulted area. The deeper they are eroded the fewer strata involved in the upthrust hence the narrower the belt of upthrust strata (Fig. 2). Between the upthrust strata are belts 1 to 3 miles wide of undisturbed strata (Fig. 3).

Crustal shortening across the faulted belt south of the mountains varies greatly but is estimated to attain a maximum of approximately 3 miles in the southeastern segment of the belt. A conservative estimate of the corresponding reduction in area is 150 square miles. In the faulted area north of the mountains the crustal shortening is probably only half as great.

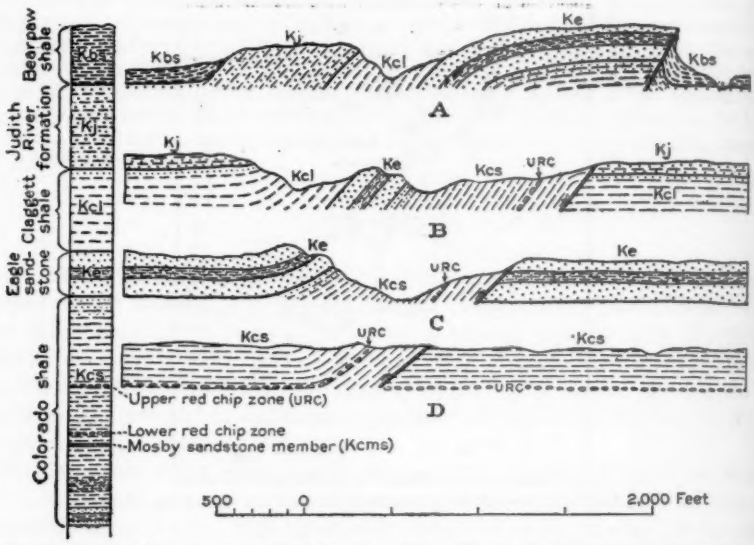

FIGURE 2.—*Cross sections showing typical structure of the upthrust belts at different erosional levels*

A, *B*, and *C* are characteristic of many faults lying respectively southeast, south, and southwest of the Bearpaw Mountains; *D*, of most of the faults adjacent to the Highwood Mountains.

GEOLOGY OF THE BEARPAW MOUNTAINS

The Bearpaw Mountains occupy an area about 35 miles in diameter and consist of an erosional remnant of Tertiary volcanic rocks overlying an eroded arch of Cretaceous and early Tertiary strata. The crest of this arch is exposed in a belt 4 to 6 miles wide which trends east through the central part of the mountains. Colorado shale and numerous stocks and laccoliths, with a core of composite dikes and sills, constitute the country rock of this belt.[1]

The volcanic rocks on the border of this belt rise to heights of 6000 to 7000 feet above sea level and 2000 to 3000 feet above the adjacent plains. The outer margin of the mountains is marked by down-faulted extrusive rocks which are tilted moun-

[1] The writer spent 3 months during the summer of 1923 making a reconnaissance map of the sedimentary and igneous rocks of the Bearpaw Mountains. The igneous geology of the western part of the Bearpaw Mountains was studied in detail by W. T. Pecora in the summers of 1937, 1938, and 1939. The results of his studies were presented as a doctor's dissertation at Harvard University in 1940. Similar studies of the igneous geology of the eastern part of the mountains were made by Bernard Fisher in the summers of 1940 and 1941 and will be presented as a doctor's dissertation at Harvard in 1946. A joint paper on the geologic history of the Bearpaw Mountains was presented by these geologists at the Boston meeting of the Geological Society of America in 1941.

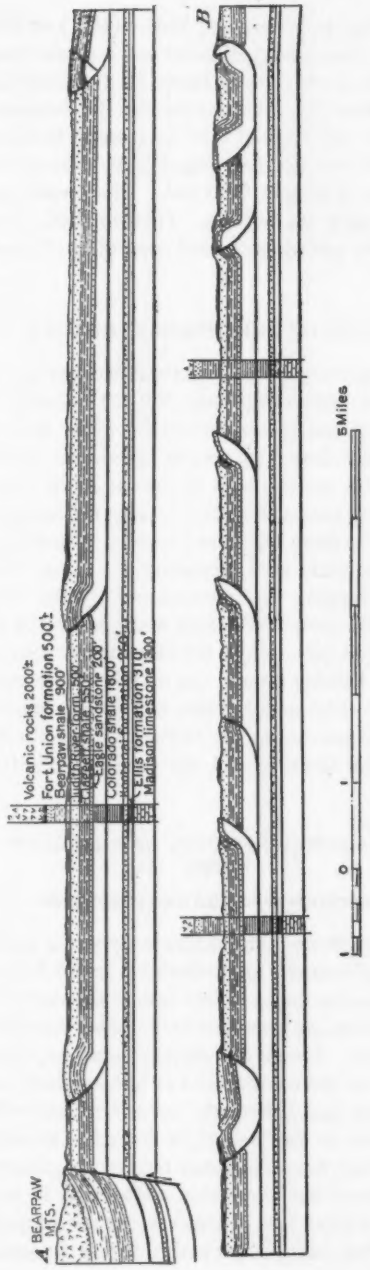

FIGURE 3.—*Cross section along line* A-B *of Plate 1*

Showing thrust faults and their probable mergence at depth with a surface of slip

tainward and rest, according to Pecora and Fisher (1941) on sedimentary rocks of Fort Union age. Between these marginal faults and the inner margin of the volcanic rocks the sedimentary floor is not exposed except in the Riedel area in the southern part of the mountains, where the volcanics rest on the Bearpaw shale and Judith River formation. Neither the original nor the present thickness of the volcanic rocks can be estimated without detailed mapping, but the volcanic pile may have had a maximum thickness of around 5000 feet. The present outer margin of the volcanics is determined largely by faulting. Their original extent on the adjacent plains is unknown, but they probably covered most of the thrust-faulted belt south of the mountains.

RESUMÉ OF THE GEOLOGIC HISTORY

The chronologic relations between the structural features and the igneous history of the region have not been fully established, but the following sequence of events appears probable: A doming and subsequent erosion of the sedimentary rocks in the present mountain area was followed by periods of volcanic activity during which a great pile of extrusive rocks accumulated in the mountain area and extended for several miles in the adjacent plains (Fig. 4). During the latter part of the volcanic activity, probably during its more explosive phases, a plainsward sliding took place down the flanks of the mountain arch, producing the thrust faults in the adjacent plains. This plainsward slipping was accompanied by the formation of a rift in the volcanic rocks along the crest of the arch and possibly by the intrusion of the dikes common to the central belt. Near the close of the volcanic activity, or possibly after it, the normal faulting around the margin of the mountains took place. The volcanics overlie the Fort Union formation, but the late Miocene or early Pliocene gravel benches around the mountains are neither faulted nor intruded by igneous rocks. Therefore both the faulting and the igneous activity occurred in early Tertiary time.

HYPOTHESIS OF PLAINSWARD SLIDING OF A SHALLOW PART OF THE CRUST

CONDITIONS FAVORING THE HYPOTHESIS

Plainsward inclination of strata.—The sedimentary strata underlying the volcanic material in the Bearpaw Mountains are arched in a broad fold trending east. The inclination of the strata on the flanks of this fold at the time of the thrust faulting is not known, as the structure has been modified by post-volcanic subsidence around the margin of the mountain. Rough calculations, however, based upon the present height of the mountain arch and assumptions as to how much it was lowered by the marginal subsidence suggest that it formerly had a structural relief of approximately 5000 feet and that the strata on the flank of the dome had an average plainsward dip of around 3°. This dip may have decreased toward the plains and merged with a slight plainsward tilt beyond the mountains. However, in making a quantitative estimate of the force in the tilted part of the crust for the purpose of determining its ability to produce the thrust faults in the plains, it may be assumed that the tilting

FIGURE 4.—*Sketches depicting possible stages in the development of the structure of the Bearpaw Mountains and adjacent plains*

was confined to the present mountain area; that is, the strata dipped outward at 3° for a distance of approximately 20 miles, beyond which they were flat.

Presence of slippery bentonite beds.—Frictional resistance to slipping must have been slight to allow strata to slide down such a gentle slope and to thrust fault the adjacent flat-lying strata. The bentonite beds in the upper part of the Colorado shale, however, where the slipping presumably took place, may well have furnished a surface of slip with a low coefficient of friction, as wet bentonite is very slippery. It is probable that a bentonite bed will become wet where it is in contact with porous sandstones saturated with water. Many bentonite beds and thin layers of sandstone occur in the upper half of the Colorado shale where the slipping is thought to have occurred, and probably one or more of these bentonite beds could have absorbed sufficient water to have formed a slippery surface.

Overloading by volcanic rocks.—Volcanic material accumulated in the mountains on a large scale previous to the thrust faulting in the plains. The weight of this volcanic mass, which may have attained a thickness of 5000 feet in the central part of the mountains, is thought to have been the main cause of the predicated plainsward sliding of the crust.

Weakness of strata acting as buttress to the mountain arch.—In ordinary landslides, owing to the comparatively steep slopes on which they are formed, the surface of slip emerges from the ground allowing the landslide mass to override its foreground without the latter offering any appreciable opposition to the movement. Due to the slight angle on which the Bearpaw sliding is assumed to have occurred it was difficult for the mountain mass to override its foreground. This, however, was offset by the weakness of the strata in the foreground which would readily buckle and thereby allow the mountain mass to move plainsward.

Earth tremors accompanying volcanic eruptions.—Most of the volcanic material of the Bearpaw mountains is of the central or explosive type, the ejection of which was undoubtedly accompanied by numerous earthquakes, for all explosive eruptions known to history have been accompanied by earth tremors. Six thousand shocks were reported by Jaggar (1924) in Hawaii during the explosive outbursts in May 1924. It is probable therefore that earth tremors were an important factor in the sliding movement. They actually may have been the main factor in initiating the slipping by overcoming statical friction by shearing off dikes that crosscut the surface of slip and by producing the ruptures along which faulting took place. Possibly after these obstructions to free movement were removed, gravity alone was sufficient to overcome kinetic friction, which may have been appreciably less than statical friction.

ANALYSIS OF THE MECHANICS INVOLVED IN THE HYPOTHESIS

General statement.—In the following mathematical analysis of the hypothesis the only factors considered are the gravitational force exerted by the tilted sedimentary crust with its load of volcanic rocks and the resistance offered to this movement by statical friction plus the force necessary to form the outermost fault. The quantitative data on which these factors are based are subject to considerable uncertainty. Moreover, certain intangible factors, such as the effect of earthquake shocks, irregu-

larities on the surface of slip, and the possibility of movement on more than one bentonite bed, are ignored. In spite of its admitted limitations, the analysis tends to show that the hypothesis is mechanically tenable. It also makes it possible to depict graphically the distribution of stress in a unit prism of the sliding crust, which is useful in explaining some of the unusual features of the thrust faults.

The analysis will be preceded by a discussion of the coefficient of statical friction.

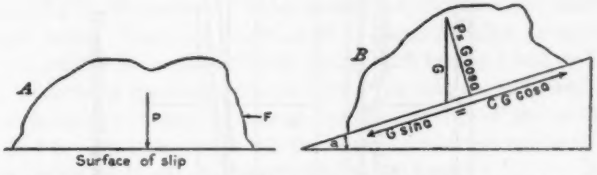

FIGURE 5.—*Diagrams showing relation under two sets of conditions, between surface of slip and forces tending to produce motion, against friction, of a body resting on it*

A. Surface of slip is horizontal. Pressure exerted thereon by object in contact with it = P. Motion begins when horizontally applied force F slightly exceeds frictional resistance. B. Surface of slip is inclined. Pressure P exerted by object = component of gravity ($G \cos a$) acting normal to surface of slip. Motion begins when component of gravity $G \sin a$ acting parallel to the surface of slip slightly exceeds frictional resistance, or $C\,G \cos a$.

Statical friction is the resistance encountered in initiating movement on a surface and can be roughly evaluated. After motion begins, it is opposed by kinetic friction, which may be much less than statical friction. Kinetic friction is indeterminate especially when an enormous mass moves on a slippery surface such as is under consideration. The diagram presented in Figure 5 will help to explain how the coefficient of statical friction is derived.

To start an object sliding on a horizontal surface, a horizontal force slightly in excess of the frictional resistance must be applied. The ratio of this force (F) to the pressure (P) which the object exerts on the underlying surface in a direction normal to it is the coefficient of statical friction (C), hence $C = \frac{F}{P}$, or, as it is usually stated, $F = CP$. If the plane is tilted sufficiently the object will slide of itself. The angle that the surface of slip makes with the horizontal when slipping begins is called the angle of repose. If G equals the weight of the tilted object and a is the angle of repose, then the weight of the object is resolved into the components $G \sin a$ acting parallel to the inclined plane and $G \cos a$ acting perpendicular to the plane. Then $G \sin a$ is the force causing movement, and $CG \cos a$ is the frictional resistance opposing it. Since these two forces are approximately equal, it is evident that the coefficient of friction involved in sliding an object on a given surface is equal to the tangent of the angle of repose, for if $G \sin a = CG \cos a$, then $C = \frac{G \sin a}{G \cos a}$ or $\tan a$.

Dimensions and weight of unit prism of sliding crust.—A unit prism 1 foot wide and with the length and height of a cross section of the sliding part of the crust will now be considered. Figure 6 represents such a unit prism. The tilted part of the sliding

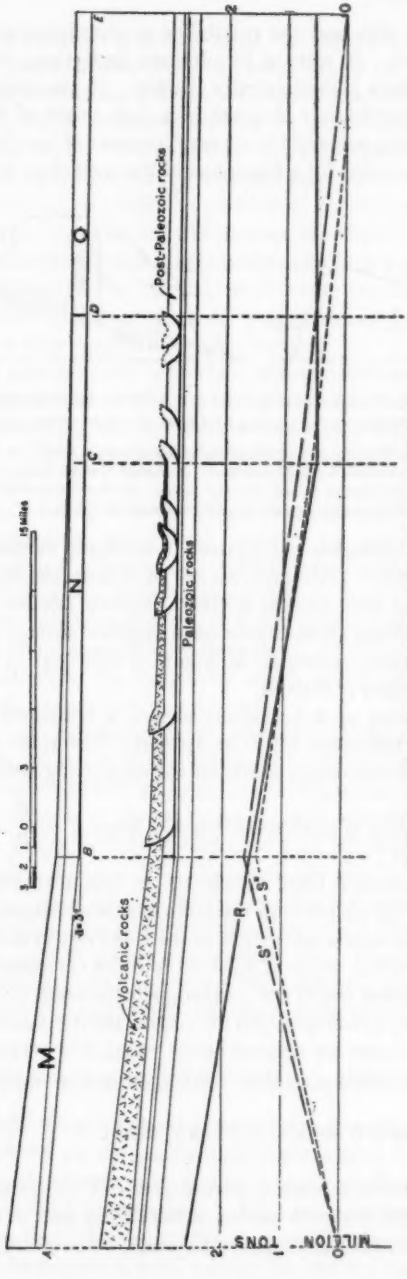

FIGURE 6.—*Unit prism of sliding part of the crust (MNO)*

With graphs showing its crushing strength (*R*) and the distribution of stress *S'* and *S''* at different stages in its development

crust is represented by M; the horizontal part, wherein faulting occurred, by N; and a more distant part, only elastically deformed, by O. The surface of slip, $ABCDE$, is assumed to be that of a bedding plane approximately 750 feet below the top of the Colorado shale. This plane underlies the volcanic rocks at about 3500 feet in the outer part of the mountains, and 1500 feet in the central part. Assuming a former thickness of 2500 feet for the volcanic rocks in the outer part of the mountains and 5000 feet in the central part, a total average thickness of 6250 feet is obtained for M, 60 per cent of which consists of volcanic rocks and 40 per cent of sedimentary rocks. Since the length of M is about 20 miles, or approximately 100,000 feet, its volume is therefore 100,000 feet by 6250 feet by 1 foot, or 625,000,000 cubic feet. Assuming a specific gravity of 2.9 for the volcanic rocks and 2.7 for the underlying sedimentary rocks, an average specific gravity of the entire mass is approximately 2.8. Each cubic foot will then have an average weight of 0.088 ton, and the total weight of M will be roughly 55,000,000 tons.

The sedimentary crust in N originally was probably about 3500 feet thick, and covered, between B and C, by a mantle of volcanic rocks approximately 2500 feet thick at B and gradually thinning to zero at its outer margin somewhere near C. The average thickness of N, therefore, was approximately 4750 feet, about four-fifths of which was sedimentary rock and one-fifth volcanic rock. The length of N was about 29 miles or approximately 150,000 feet. Its volume, therefore, was 150,000 feet by 4750 feet by 1 foot, or approximately 712,500,000 cubic feet. Assuming a specific gravity of 2.9 for the volcanic rock and 2.5 for the sedimentary rocks, which are less indurated than the sedimentary rocks in the mountain area, an average specific gravity of 2.58 is obtained for N. Each cubic foot will therefore have an average weight of approximately 0.080 ton, and the total weight of N will be around 58,000,000 tons.

Quantitative estimate of forces.—If slipping occurred on $ABCD$ of Figure 6, it was caused by the component of gravity of M acting parallel to AB. If this force produced the faulting in N, at its maximum, it must have been equal to, and not materially greater than, the frictional resistance encountered in M sliding on AB and N sliding on BD, plus the force necessary to produce the outermost fault at D. If G = weight of M, or 55,000,000 tons; G' = weight of N, or 58,000,000 tons; C = coefficient of statical friction; a = inclination of M to the horizontal, or 3°; R = force necessary to produce the fault at D, or 500,000 tons[2], then the component of gravity causing the slipping = $G \sin a$ and the frictional resistance on AB and $BD = CG \cos a$ and CG', respectively, and $G \sin a = CG \cos a + CG' + R$ (*1*) and

$$C = \frac{G \sin a - R}{G \cos a + G'} \quad (2).$$

Substituting in (2):

$$C = \frac{2{,}878{,}700 - 500{,}000}{54{,}924{,}650 + 58{,}000{,}000} = .0216 \quad (3)$$

[2] The value of R is assumed to be equal to the crushing strength of the rocks in the forepart of N. The shales in the section are regarded as having practically no crushing strength. The sandstones have a total thickness of approximately 1200 feet but are not well indurated and are assumed to have a crushing strength of around 6000 pounds per square inch, or 432 tons per square foot. On these assumptions the total crushing strength of the forepart of N is 1200 times 432 tons, or approximately 500,000 tons.

Assuming a value of 0.0216 for the coefficient of friction, M, if unopposed by N, would have slipped plainsward on a surface whose angle of slope had a tangent value of 0.0216 or an angle of approximately 1°. Consequently, the amount that M was tilted above this angle gave it the force to shove N forward on the surface BD and produce the thrust faults. This was evidently equal to the total force exerted by the component of gravity of M acting parallel to the surface of slip minus the frictional resistance encountered by movement on that surface. $G \sin a - CG \cos a = 1{,}692{,}000$ tons. The horizontal component of this force, effective in shoving N forward, is equal to cos a times 1,692,000 tons, or approximately 1,600,000 tons.

A graphic representation of stresses in a unit prism of the sliding mass and their relation to the crushing strength of the rocks is presented in Figure 6. Graph R of this figure represents the crushing strength of the rocks in any vertical section of the unit prism. It is based upon calculation of the total crushing strength at points A, C, and D. The total crushing strength at D, as previously estimated, is 500,000 tons. Beyond the area formerly covered by volcanic rocks the thickness of the sliding part of the crust probably varied little, and the value of R is assumed to be constant between C and E. The total crushing strength of the rocks increases regularly from the former outer margin of the volcanic rocks at C to their thickest part, in the center of the mountains, where it is roughly estimated to be five times its value at D, or 2,500,000 tons.

The theoretical distribution of the stress when it attained its maximum intensity is shown in graph S'. It is based upon a determination of the values of S', at A, B, C, D, and E. Its value at A is zero. Calculations made above give its value at B as 1,600,000 tons and at D as 500,000 tons. The graph between B and E is not a straight line because more force is required per unit distance in shoving forward the part of N laden with volcanic rocks than the part not so laden. The value of S' at C equals its value at B minus the frictional resistance between B and C, which is roughly 900,000 tons. Its value at C therefore equals 1,600,000 tons minus 900,000 tons, or 700,000 tons.

The value of S' in O, the part of the crust that is only elastically compressed, decreases from 500,000 tons at D to zero at some distant point E. The distance DE or the length of O is governed by the frictional resistance encountered in the slipping that accompanies the elastic shortening. It may be determined as follows:

Let F be the maximum force applied to O in cross section, P the weight of the rocks elastically shortened by this force, and C the coefficient of friction then, by definition,

$$C = \frac{F}{P} \tag{1}$$

$$F = PC \tag{2}$$

If d be the length and h the thickness of the rocks in O and w their weight per cubic foot (0.08 tons) then

$$P = dhw \tag{3}$$

Substituting in (2),

$$F = dhwC \tag{4}$$

$$d = \frac{F}{hwC} \tag{5}$$

$F = 500{,}000$ tons; $h = 3{,}500$ feet; $w = 0.08$ tons; $C = 0.0216$. Substituting these values in (5), $d = 82{,}600$ feet or approximately 15 miles.

GEOLOGIC FEATURES EXPLAINED BY THE HYPOTHESIS

General statement.—The above hypothesis offers a reasonable explanation of how a thrust force of sufficient intensity to produce the deformation in the plains could have originated from the mountain area and explains the shallowness of the deformation which otherwise would be inexplicable. It also accounts for the absence of volcanic rocks in the central part of the mountains, for, inasmuch as this belt occupies the crest of the mountain arch, it may be regarded as a rift produced by the sliding crust. The narrow width of the belt and the fact that it extends entirely across the mountain are not consistent with the assumption that it is primarily an erosional feature. Erosion no doubt has widened the original rift to some extent, especially at its eastern and western ends. Allowing for some widening by erosion, the rift can be assumed to have covered an area of around 200 square miles, which is roughly equivalent to the estimate of the reduction in area of the crust produced by the thrust faulting in the surrounding plains. Possibly the separation of the shallow part of the sedimentary crust beneath the volcanic rock resulted in the formation of the many east-west trending dikes that occupy the central belt.

Prevalence of thrust faulting.—Yielding of the strata in the plains by thrust faulting requires explanation. The bedded rocks of the earth's crust commonly respond to horizontal pressure by folding, with attendant thrust faulting. Why this did not take place in the area under consideration may be attributed to both the shallowness of the crust involved and the weakness of the strata. In shallow deformation such as that under consideration, the rocks involved must transmit all the forces required in bending and lifting the strata to form a fold, because beyond a slight elastic expansion the underlying rocks will not accompany or contribute to the folding. The preponderance of shales in the involved part of the crust made it incapable of forming unsupported arches. The sandstones present, although sufficiently thick and strong to control the breadth of span of any flexure that tended to form, were unable to lift the load of the rocks involved in such folds. At an early stage in the deformation therefore faults were formed, along which further deformation took place.

Uniformity in throw of the faults.—That there is no gradual decrease in the throw toward the outer margin of the faulted area is seemingly not in accord with the concept that the deformational force was transmitted outwardly from the mountain area, for evidently the force would be at its maximum near the mountains and would be reduced by frictional resistance to zero at some distant point. The throw of the faults from *B* to *D* in the cross section shown in Figure 6, which may be considered as illustrating fairly typical conditions, is roughly 700, 700, 700, 900, 1000, 1600, 1600, 1400, 1600, and 1600 feet. A graphic analysis of the growth of the stress and its relation to the crushing strength of the rocks as shown in Figure 6 makes it evident why the faulting does not decrease in intensity outward from the mountains. Graphs showing the stress at different stages in its growth will closely parallel graph S' representing its maximum development. The smallest stress that will exceed the crushing strength of the rocks in any part of the unit prism is represented by

graph *S''*. At the stage at which this stress was exerted, the crushing strength was exceeded at *C*, and a rupture was formed. Movement on this rupture plane continued until the resistance encountered in further lifting the rocks resulted in increase of stress to a point where the crushing strength of the rocks was again exceeded. This will occur to the right and left of *C* (Fig. 6), first close by and then at increasingly greater distances until the last faults are formed at *B* and *D*. If this were the order of faulting, it presumably should be indicated by the throw of the faults, the first fault to form showing the greatest throw and the last one the least. Although the fault at *C* has a throw of 1600 feet and the faults on the mountainward side of *C* show a tendency to decrease in throw away from *C*, those on the plainsward side are major faults. The greater throw of the faults on the plainsward side of *C* may be due to the fact that their growth was not retarded by the load of volcanic rocks which are assumed to have occupied the area on the mountainward side of *C*. The occurrence of a major fault at *D* still remains to be explained. It would be expected that movement on this fault would be small or not much greater than that required to restore the stress to its intensity when the rupture was formed. Instead, it has a throw of 1600 feet, and farther plainsward the strata are not disrupted by a single minor fault. The phenomenon would be explicable if some obstruction to faulting existed beyond *D*, such as is presented by the laccolithic domes on the southwest side of the Little Rocky Mountains, which causes a deflection in the trend of the faults in that area. Yet, elsewhere along the margin of the faulted area there are no such barriers, and there is no reason to assume that the crushing strength of the rocks or frictional resistance was greater outside than within the faulted area. The occurrence of a major fault at *D* and the absence of any minor faults farther plainward seems to imply that when the rocks ruptured the intensity of the stress existing at the initiation of the rupture was not restored until a fault of major throw was produced. This appears theoretically doubtful and also is not consistent with the fact that there are many faults with throws of 700 feet and less. The only explanation of the anomaly that the writer can offer is that the stress reached its maximum intensity when this fault was initiated and that thereafter the erosion of the upthrust strata along the faults prevented any further growth in the intensity of stress. The slight difference in the throw of the major faults can be attributed to slight differences in the factors governing the growth of the faults.

Spacing of faults.—The faults are separated by broad belts of undisturbed strata. In most regions of crustal deformation by horizontal pressure, the strata everywhere show in some way the effect of the deformation, either by being folded into a continuous series of anticlines and synclines or thrust-faulted. The spacing of the thrust faults under consideration is illustrated in the section shown in Figure 6. Between *C* and *D* they occur at intervals of 1 to 2 miles. Between *B* and *C* the spacing ranges between 3 and 4 miles. Although the spacing in *CD* could scarcely have been closer without crowding the disturbed belts, in *BC* there was room for other faults to have formed. Under the special conditions of low coefficient of friction and the gradual thinning of the extrusive rocks from *B* to *C* the faults most widely spaced may have been formed at about the same time in the development of the stress and their spacing was more or less fortuitous. After the formation of these faults, erosion of

the upthrust strata and the attendant movement on the fault planes may have prevented the stress from growing in intensity and consequently no other faults would form.

GEOLOGIC FEATURES NOT EXPLAINED BY THE HYPOTHESIS

There are admittedly structural features in the faulted area for which the hypothesis offers no ready explanation. Among these is the occurrence of several faults which are not circumferential or radial to the mountains and are therefore seemingly due neither to an outward thrust movement from the mountains nor to the shearing of the crust in such a movement. Tilted strata along the apparent shear faults also show the same type of structure as the strata in the upthrust belts bordering the circumferential faults. Moreover, beyond the marginal fault on the southeast side of the faulted area the flat-lying Cretaceous formations lie 150 feet lower than they do within the belts of undisturbed strata between the thrust faults (Fig. 3). This feature can be observed in the gorge that the Missouri River has cut across the southeast margin of the faulted area. Because of inadequate exposures it is not known whether a similar difference in elevation of the strata occurs elsewhere along the outermost fault.

The above features suggest an expansion of the crust in all directions which seemingly does not fit in with the hypothesis of a one-sided movement originating within the mountains. Possibly, however, this expansion may be attributed to dilatancy—the property of granular masses to expand in bulk with change in shape. The consequences of dilatancy by deformation are said by Mead (1925) to have far-reaching geological significance. However, the evidence of the increase in bulk of the strata may point to some as yet unrecognized principle which may be the chief factor in the formation of thrust faults and other structural features attributed to horizontal compression.

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McCARTYS BASALT FLOW, VALENCIA COUNTY, NEW MEXICO

BY ROBERT L. NICHOLS

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ABSTRACT

The McCartys flow is approximately 30 miles long, in places several miles wide, approximately 1.7 cubic miles in volume, and it has an area of about 119 square miles. The flow is in the main unweathered, uneroded, and barren. Pueblo I potsherds have been found in a valley fill which is probably buried by the flow, indicating that extrusion occurred in the last 1200 years.

Minor features such as grooved lava, squeeze-ups, ropes, spatter cones, tree molds, cracks, banded lava, and cavities are present. The extrusion of the flow impeded the Rio San Jose and diverted drainage from it into the Rio Salado and Little Colorado. The flow moved by the flow-unit and single-unit mechanisms, and its velocity varied greatly from place to place.

Near the terminus of the flow are about 100 collapse depressions. Analysis of the probable strength of the roofs of lava tunnels indicates that they owe their origin not to the collapse of one large tunnel but to the partial collapse of several small tunnels. In some cases collapse occurred while liquid lava was still present and when the under side of the crust was still soft and plastic.

On the last mile of the flow pressure ridges are very common. They are as much as 1200 feet long, 25 feet high, and 100 feet wide. In transverse section they resemble the gable of a house. In general they parallel the flow and are close to its margin. They formed when liquid lava was beneath the crust. It is suggested that the collapse of a dome between 27 and 38 feet high, due to withdrawal of lava from beneath, resulted in compressional forces which buckled the crust and produced the pressure ridges.

INTRODUCTION

The McCartys basalt (Fig. 1) is a relatively fresh pahoehoe flow whose striking features have long excited interest in travelers on the Santa Fe Railroad between Albuquerque and Gallup, N. M. The flow originated southwest of the Rio San José and occupies the valley from a point about 7 miles east of Grants to the little Indian town of McCartys, a distance of about 6 miles. From U. S. Highway 66 much of this flow is visible as the highway passes across and beside it. The writer spent parts of the field seasons of 1933, 1934, and 1935 making a detailed study of the lower 6 miles of the flow. In 1938 he returned for a period of 3 days to obtain further details. The present paper resulted from this study.

ACKNOWLEDGMENTS

The writer is greatly indebted to Professor Kirk Bryan of Harvard University, under whose supervision the study was undertaken and completed. Professor Bryan critically read the manuscript, spent long hours with the writer in discussing the problems involved, contributed numerous ideas, and spent several days in the field with the author.

Warm thanks are also due Professor Esper S. Larsen for help in the laboratory and for many suggestions. Professor R. A. Daly's knowledge of the volcanologic literature was of great assistance, and discussion with Professor A. C. Lane on the rate of cooling of lava and on grain size was also of much help.

Thanks are due Mr. Gordon Taylor, Mr. Lawrence G. Ropes, Professor F. N. Weaver, Dr. Laurence LaForge, and Dr. Richard Tousey, who helped the writer apply engineering principles to the solution of the problems considered.

FIGURE 1.—*Index map*
Location of McCartys flow shown by diagonal lines.

To Karl Benedict, Robert S. Folsom, David Rose, Ralph Fellows, Edmund Shaw, Charles E. Stearns, and Charles E. Baker, of Tufts College, thanks are due for able field assistance.

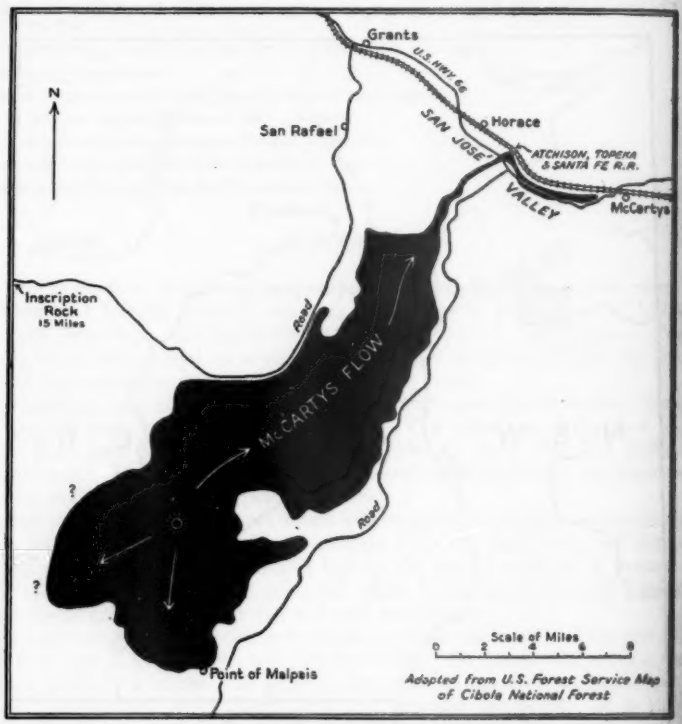

FIGURE 2.—*Approximate outline of McCartys flow*

GENERAL ASPECTS OF THE FLOW

The source of the McCartys flow is in a broad valley between mesas about 20 to 25 miles southwest of the point where the flow is crossed by U. S. Highway 66 (Fig. 2). Here, at a point about 1000 feet higher than the San José Valley, a small cinder cone rises approximately 50 feet above a broad lava cone (Pl. 1, fig. 1). From the lava cone the lava flowed northeastward to the San José Valley and then eastward down the valley, finally stopping a few hundred yards west of McCartys. Lava also flowed from this cone southwestward about 6 miles. Near its source the flow is several miles wide; it gets progressively narrower, however, until it is less than 200 feet wide where it enters the San José Valley. The average width in the valley is about 1000 feet. Figure 2 of Plate 1 shows the flow 1½ miles west of McCartys where it is about 1500 feet wide and fills the valley. Near the source gush after gush built up a lava cone at least 200 or 300 feet thick. The approximate thickness

FIGURE 1. SOURCE OF McCARTYS FLOW

FIGURE 2. FLOW 1½ MILES WEST OF McCARTYS

FIGURE 1. NARROW THREAD OF LAVA WHERE McCARTYS FLOW WAS CONFINED BY CANYON

FIGURE 2. POND IN ONE OF THE COLLAPSE DEPRESSIONS OF THE McCARTYS FLOW

McCARTYS FLOW CONFINED BY CANYON AND POND IN COLLAPSE DEPRESSION

of the last 6 miles of the McCartys flow may be obtained by studying the depth of the larger cracks, the edge of the flow, the depth of the collapse depressions, the edge of the flow around a kipuka, and by a comparison with the Laguna flow (Nichols, 1934), which is similar petrographically but eroded so that its thickness can be measured. Measurements around the kipuka (Fig. 3) indicate that the flow in this

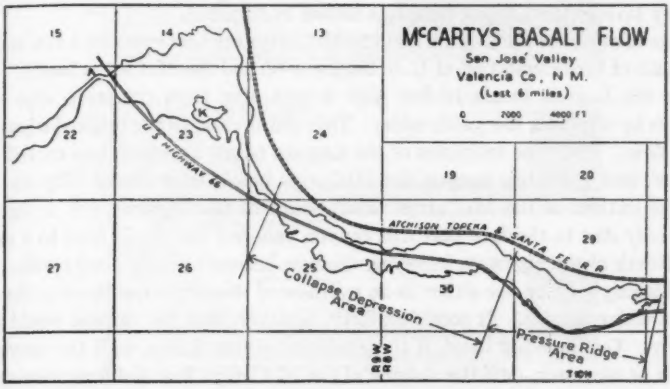

FIGURE 3.—*Last six miles of McCartys flow*

Area marked *K* is a kipuka. Position of collapse depression area and pressure ridge area indicated. Location of Figure 4 indicated by *A*.

area is about 15 feet thick; however, where the flow narrows it is probably as much as 50 feet thick. The last 6 miles of the flow is probably 30 to 40 feet thick. The average thickness of the Mull and Deccan flows is approximately 50 feet (Anderson, 1924; Fermor, 1925). The flow covers approximately 119 square miles, and if 75 feet is taken as the average thickness of the whole mass the volume is 1.7 cubic miles, which is much greater than that of the average basaltic flow in western United States. It has nevertheless only one-sixth the volume of the Veidivatnahraun basaltic flow of Iceland (Daly, 1933, p. 139). This flow, however, was erupted from a fissure 150 kilometers long, whereas the McCartys flow was erupted from a single vent.

Although the McCartys flow is not eroded and its floor has never been seen, it must rest either on Mesozoic sandstone and shale in which the valley is eroded, on the Laguna flow, or on the stream deposits which floored the valley at the time of extrusion.

AGE OF THE FLOW

The Rio San José lies in a valley a quarter of a mile to 2 miles wide which has been incised to a depth as great as 1200 feet in a basalt-covered erosion surface which Bryan and McCann (1936; 1938) consider to be late Pliocene or Pleistocene. As the McCartys basalt occupies the bottom of this valley, the flow cannot be older

than late Pleistocene; its fresh appearance and the retention of surface features herein described prove that it is Recent and probably about 1000 years old.

The McCartys basalt is younger than the Laguna basalt. Pressure ridges of the Laguna project through the McCartys as miniature steptoes. At several places the McCartys basalt filled collapse depressions of the Laguna basalt. The Laguna is in places eroded, whereas the McCartys is uneroded. A kipuka in the McCartys, floored by lava of the Laguna basalt, is shown in Figure 3.

The age relations of the Laguna and the McCartys are best revealed a few hundred yards south of the intersection of U. S. Highway 66 and the McCartys basalt. Here a cliff in the Laguna basalt 10 feet high is separated from the steep edge of the McCartys by a moat a few yards wide. This cliff is close to the original edge of the Laguna flow. After the extrusion of the Laguna basalt a canyon was carved along its border, and down this canyon the McCartys basalt later flowed (Fig. 4). The threadlike outline of the McCartys basalt south of the highway (Pl. 2, fig. 1) is undoubtedly due to the fact that this canyon confined the liquid lava to a narrow path. North of the highway, however, the flow becomes nearly a mile wide. This rapid widening may be due either to an overflow of the canyon walls or to the existence of a wider canyon. It seems unlikely, however, that the canyon would widen so rapidly. On the other hand, if the gradient became flatter, or if the canyon got shallower or narrower, or if the volume of the McCartys flow suddenly increased, it might well have overflowed the walls of the canyon.

The McCartys flow over most of its area is as fresh and unweathered as the historic flows of the Hawaiian Islands, Vesuvius, and other areas. Weathered lava is rare. Below the unweathered glassy surface of the flow, however, the lava is in places discolored by limonite. On the last 6 miles of the flow, widely scattered plants such as lichen, cactus, sage, and small juniper grow. In many places the south side of a pressure ridge is black and barren, while the shady north side is covered by lichen. Where the crust is fragmented (Nichols, 1938, p. 601–603) the surface is usually bare, but in places the north side of the fragments is covered with lichen.

Near the source of the basalt, more than 1000 feet higher than that part of the San José Valley occupied by the McCartys flow, each margin of the flow has a belt of vegetation less than a mile wide. In these marginal belts, the yellow pine is as much as 3 feet in diameter. However, with increasing distance from the edge of the flow the pines gradually decrease in size and number, and the central portion of the basalt is bare. These belts of vegetation grow on wind-borne soil. The wind picks up dust from the area marginal to the flow, and the dust cover thins with increasing distance from the edge of the flow. The relatively heavy vegetation near the source and the barreness of the terminus are apparently controlled by the difference in rainfall because of difference in elevation. Similar differences in vegetation are also found on Hawaiian flows. Yellow pine of this size on the flow indicates that it must be several hundred years old.

The Acoma Indians, on whose reservation a small part of the McCartys flow is found, have a legend that the flow covers land their ancestors tilled. This legend (Darton, 1915), if credible, dates the eruption of the basalt as pre-Spanish but later than about 700 A.D.—the beginning of the pueblo civilization.

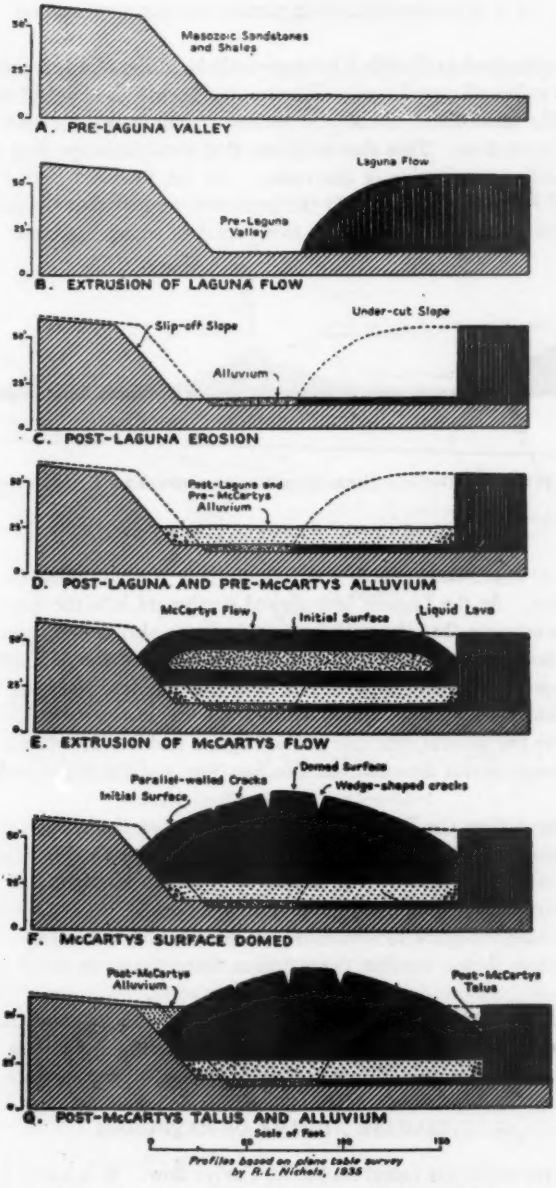

FIGURE 4.—*Relation between Laguna and McCartys flows*

These cross sections were made at the narrowest part of the flow shown in Figure 1 of Plate 2. Studies farther down the valley indicate that following the erosion of the Laguna a valley fill was deposited which was later buried by the McCartys.

Potsherds identified as Pueblo I have recently been found 4 feet below the top of the youngest valley fill near Laguna (Reiche, 1936), which is about 17 miles down the valley from the terminus of the McCartys flow. This valley fill is probably buried by the McCartys flow. This also indicates that the McCartys flow was extruded during the Indian occupation of the valley. As the Pueblo I period was between 700 A.D. and 900 A.D. (Hack, 1941) the potsherds suggest that the extrusion of the McCartys flow occurred either during this period or somewhat later.

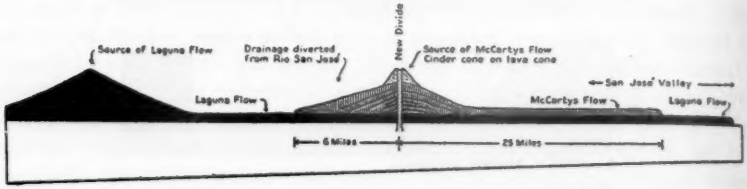

FIGURE 5.—*Drainage diversion resulting from extrusion of McCartys flow*

DRAINAGE CHANGES EFFECTED BY THE FLOW

The cone or cones from which the Laguna flow was extruded lie south of the McCartys flow. As the Laguna lava flowed northward into the San José, it seems reasonable to suppose that the post-Laguna drainage also flowed from the source of the Laguna northward. However, with the extrusion of the McCartys flow some of the drainage was apparently diverted from the San José southeastward into the Rio Salado and possibly southwestward into the Little Colorado River. Figure 5 not only shows the general relations of the Laguna and McCartys, but it also shows how the drainage divide between the Rio San José and the Rio Salado was shifted northward.

At the present time the Rio San José, in that part of the valley occupied by the McCartys flow, is broken up into several branches which flow beneath, on top of, or marginal to the McCartys. The branch which flows on top of the McCartys for a short distance has deposited some alluvium although it has not accomplished any significant erosion. Figure 6 shows four branches of the Rio San José at the terminus of the McCartys flow. Farther downstream these branches unite into one main stream.

Shallow lakes marginal to the flow, quite common after a heavy rainfall, indicate that the flow impedes drainage. Moreover, as is shown by Figure 2 of Plate 2, small ponds are at such times found in many of the collapse depressions of the flow.

CLIMATE AT TIME OF EXTRUSION

Only one tree mold was found on the McCartys flow. It is about 7 feet long and indicates that the tree was about 3 feet in diameter. Neither the tree nor its mold had been transported any great distance, as the mold was found only a few miles from the source of the flow. Only one traverse of the flow was made in this area, so that similar molds could probably be found. Such molds indicate a climate, at

the time of the extrusion of the McCartys flow, similar to the existing one or to one of greater rainfall.

The last 6 miles of the flow was studied in great detail, and no tree molds were found. Moreover, there is nothing to indicate that conditions existed which would

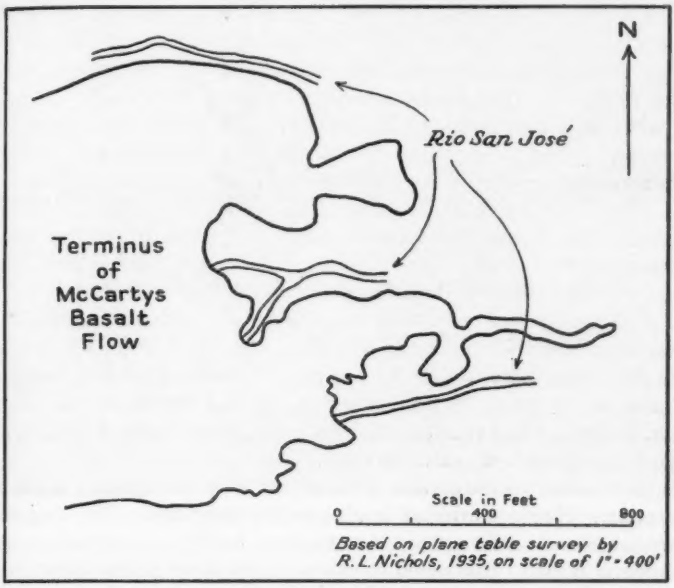

FIGURE 6.—*Terminus of McCartys flow*

Four branches of Rio San José at the terminus of the McCartys flow and also the two terminal tongues, one of which projects from the main mass of the flow 1000 feet down the valley.

have prevented the formation of tree molds. Hence, when the McCartys flow ran into this part of the valley the trees were not large or numerous enough to make tree molds. Today, juniper is abundant. If juniper were engulfed in molten lava it would probably be entirely consumed. Thus the climate in this part of the valley at the time the flow was extruded was either similar to that of the present or drier. As no significant diastrophic or topographic changes have taken place since the extrusion of the flow, the difference in climate between the area at its source and at its terminus must be about the same today as before the extrusion. It is concluded, therefore, that the climate at the time of the extrusion of the flow was similar to that of the present in all parts of the area.

FLOW MECHANISM

At the end of the flow (Fig. 6) are two tongues, one of which projects 1000 feet down the valley. A comparison of the surface gradient of the flow at this point

with a reasonable pre-McCartys gradient indicates that these tongues are not so thick as the main flow. They are therefore unburied flow units (Nichols, 1936), and their existence suggests that the flow, at least near its terminus, was advancing by the flow-unit mechanism.

The walls of a few of the larger cracks reveal miniature flow units which are a few inches to a few feet thick and can be traced for scores of feet along the walls. In places, only one miniature flow unit will be found on the main body of the flow; in other places, several such units exist, one upon the other. The existence of these flow units is indicated: (1) by the sharp contact between them; (2) by the columnar jointing which varies in size from unit to unit; and (3) by the ropy surfaces which were formed on top of the units and were buried by succeeding units. These flow units are much smaller than those described from the Laguna and Suwanee flows (Nichols, 1936). Undoubtedly the units broke out not from the sides or front of the advancing lava tongue, but from its surface. Apparently the flow thickened intermittently by the extrusion onto its surface of these miniature flow units. Richey (1924) has described similar features from the lavas of the island of Mull, and Jones (1937, p. 873) has observed them in Hawaiian flows in the process of formation.

The miniature flow units of the McCartys flow were found south of U. S. Highway 66 where the flow is narrow. There was a greater tendency for the crust to break here because the surges of the lava were stronger than where the flow was wider. Moreover, the crust would tend to be thinner here than elsewhere because of a more continuous flow of hot lava below the crust.

Ropes are common on the surface of the flow. They vary in size, shape, and in other respects. They are oriented in all possible directions. This suggests that they were not formed on the surface of molten lava flowing as a single unit, because all the currents in such a flow would be moving in the same general direction. The ropes were formed on flow units which at intervals poured from vents and cracks on the surface of the flow, each moving in various directions.

A short distance north of the point where U. S. Highway 66 crosses the flow there are several excellent examples of flow units. These flow units were not found by a study of the cross section of the flow but by an analysis of surface features.

That the flow was running with maximum velocity where it was confined by the canyon cut in the Laguna basalt (Pl. 2, fig. 1) is indicated by its greater thickness (Nichols, 1939c, p. 301) at this point and by the fact that the pre-McCartys gradient was at least as steep here as elsewhere. Because of this rapid movement, it seems likely that the flow was moving here by the single-unit mechanism.

The mechanism of flow was therefore complicated—advancing by the single-unit mechanism at one point, by the flow-unit mechanism at another; at one place slowly, at another rapidly.

MINOR FEATURES

CAVITIES

Cavities within a few feet from the surface of the McCartys flow are common. The largest observed are as much as 3 feet long, 5 inches deep, and often 3 feet or

FIGURE 1. CAVITIES IN LAVA SEEN ON THE WALLS OF A CRACK

FIGURE 2. A SPATTER CONE SIX MILES FROM THE TERMINUS OF THE MCCARTYS FLOW

CAVITIES AND SPATTER CONE

FIGURE 1. AERIAL PHOTOGRAPH SHOWING PART OF COLLAPSE DEPRESSION AREA OF McCARTYS FLOW
Many depressions are filled with water. Santa Fe Railroad on left of flow, Rio San José on right.

FIGURE 2. COLLAPSE DEPRESSION FLOORED WITH ALLUVIUM AND VEGETATION
Man standing on right-hand edge of depression gives scale.

COLLAPSE DEPRESSION AREA AND DEPRESSION FLOORED WITH VEGETATION

more below the surface. Most of these cavities have no surface expression. However, as shown by Figure 1 of Plate 3, some of the larger cavities are associated with surface blisters. The hat shown in Figure 1 of Plate 3 is resting on such a blister. It is about 5 feet in diameter and raised about 1 foot above the general level of the flow. The largest cavities are associated with the largest blisters.

Although it is patent that many cavities and gashes in the flow are due to the accumulation of gas bubbles, the author is not certain whether these cavities are due to this mechanism, to successive withdrawals of lava from bulbous squeeze-ups, or to buckling of the crust by compressional forces.

If these cavities are due to the accumulation of small gas bubbles beneath the crust, they are similar to the jagged gashes found in the Keweenawan flows described by Butler and Burbank (1929, p. 29) and to cavities in lava described by Nichols (1936, p. 619). Their formation by this mechanism can be explained as follows. If the crust of a flow is thickening slowly during a period of rapid vesiculation, one long continuous gas cavity will be formed wherever the bottom of the crust is flat. However, if the bottom is not flat, which will almost always be the case, the gas bubbles will accumulate most rapidly where the bottom is domed. In this way, unconnected cavities of a general lenticular shape, one under each dome, may be formed.

Broad swells or flat domes 10 feet across and a few inches high are found on the McCartys flow. Similar domes have been described and photographed by Anderson (1903) from the flows of Iceland. If the crust of somewhat smaller domes is sufficiently plastic and thin it will be slowly lifted, and blisters similar to those in the figure will be formed by the gas pressure. A period of rapid thickening of the crust with slow vesiculation will stop the growth of these gas cavities. After the crust has been thus thickened, conditions of rapid vesiculation and slow thickening may again obtain, and if the bottom of the crust is still domed at this point another group of gas cavities may be formed below the first. These lower cavities may also increase the size and height of the blisters. In this way superposed cavities associated with surface blisters may be formed. Such a mechanism was envisaged by Butler and Burbank (1929, p. 28) to explain the successive gashes or cavities in the Keweenawan flows. The upper cavities are older than the lower. The size and shape of the cavities depend upon the size and shape of the initial domes, the rapidity with which the crust thickened, and the rate of vesiculation. During the period of formation of these cavities the lava must have had low viscosity; otherwise the vesicles could not have migrated and coalesced.

Periods of rapid vesiculation with slow thickening of the crust followed by rapid thickening with slow vesiculation may be brought about if the forward progress of the flow is irregular rather than uniform. This condition will be satisfied if the flow is advancing by the flow-unit mechanism (Nichols, 1936) or by the single-unit mechanism if the lava was emitted from the vent with varying speeds. With such a flow mechanism, the lava would move through the carapace which surrounds it with varying velocity. During a period of rapid movement the crust would thicken slowly, and rapid vesiculation might take place, whereas during slow movement of the lava the crust would thicken rapidly, and vesiculation might also be slow.

The bottom of a gas cavity formed in this way would be flat if the lava were fluid. If the lava were viscous the cavity might be convex or concave downward, depending upon how rapidly and at what point the bubbles were added to the growing gas cavity. The top of such a gas cavity would in the main be convex upward.

These cavities imperceptibly grade in size and shape into those which are clearly due to the coalescing of vesicles, thus pointing to a formation by accumulation of gas bubbles. On the other hand, the position of the larger cavities indicates that the crust was more than a foot thick when they were formed. It seems doubtful that such a small quantity of gas as would have been trapped in these cavities could blister such a thick crust.

An alternative explanation for these larger cavities is that they are due to the withdrawal of lava from structures similar to bulbous squeeze-ups (Nichols, 1939b). These bulbous squeeze-ups are often hollow, due to a withdrawal of lava from them. According to this idea the hat in Figure 1 of Plate 3, is resting on a bulbous squeeze-up with similar structures on each side. Immediately after the extrusion of each of these supposed squeeze-ups a crust forms around them. If after the crusts of the squeeze-ups have formed and before the feeding vent is closed the hydrostatic pressure of the liquid lava is decreased, either due to the formation of a flow unit downstream or to a reduction in the quantity of lava reaching the liquid thread from upstream, lava may be drained from them. If the crust of the squeeze-up is strong enough to support itself a cavity will be formed. Progressive solidification followed by renewed drainage of lava will form additional cavities. Finally the feeding vents will be closed by solidification.

Although cavities in adjacent bulbous squeeze-ups may be formed contemporaneously, they are not necessarily the same size or shape. The ease with which the lava can be withdrawn from a squeeze-up and the size and shape of the squeeze-ups will determine the size and shape of the cavities. Figure 7 illustrates the formation of these cavities by this mechanism. A cross section through the flow immediately after the extrusion of two bulbous squeeze-ups is shown by A of Figure 7, while B shows the squeeze-ups after a crust has formed around them. C shows how a withdrawal of lava due to a decrease in the hydrostatic pressure of the liquid lava below forms a cavity in each squeeze-up. Further solidification and withdrawal of lava result in the formation of two additional cavities as shown by D and E.

The bottom of a cavity formed in this way would tend to be flat if the lava were fluid and would be concave upward with the low point near the vent if the lava were viscous. The top might be irregular due to the formation of lava stalactites, but in the main it would be convex upward.

The feeding tubes required by this mechanism have not been found. The lack of feeding tubes, however, is not fatal to the theory as the observable cross section might not have intersected the tubes, and, moreover, the material of the tubes might be so similar to the material of the floor that differentiation would be impossible.

A third explanation for these cavities is that they are due to a buckling of the crust by compressional forces, the cavities representing the space between the crust and the more or less liquid substratum. However, if this mechanism had taken place one would expect a greater abundance of tension cracks and more fragmentation of the crust.

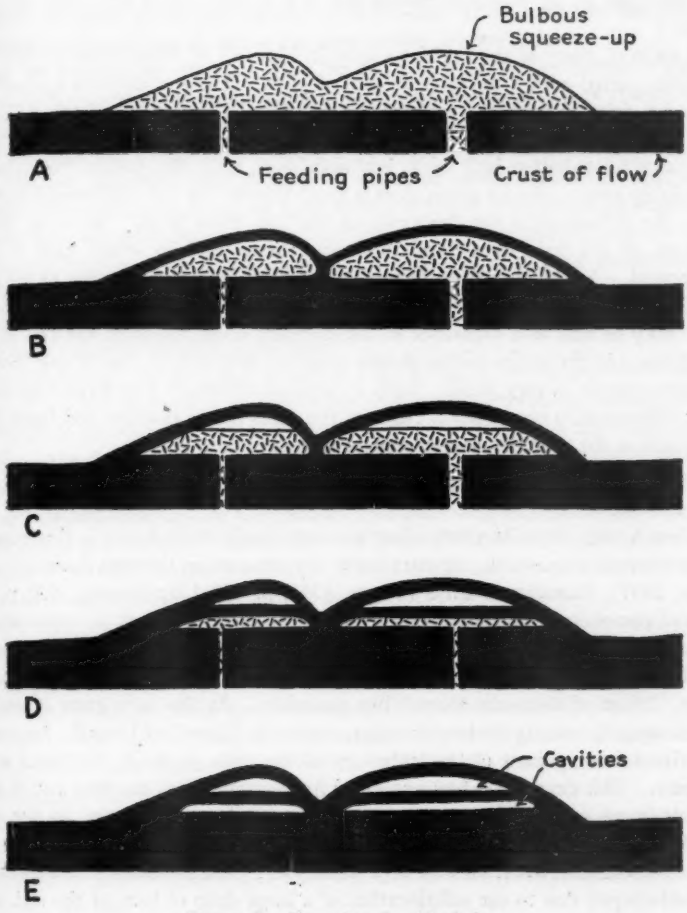

FIGURE 7.—*Formation of cavities in lava*

Diagrammatic sketches illustrating the formation of the cavities by successive withdrawals of lava from bulbous squeeze-ups.

It may well be that combined gas pressure, withdrawal of lava, and even buckling formed these cavities.

Schollendomes, low domical hills on pahoehoe flows, may be as much as 100 feet long and 15 feet high. Many are hollow. The controversy with regard to their origin is somewhat similar to the problem of the origin of the cavities under discussion. Many vulcanologists believe that they are due to the accumulation of gas beneath the crust (Sapper, 1927; Perret, 1913, p. 154). On the other hand Daly (1933, p. 155) and Jaggar (1931) believe that they are due to the hydrostatic pressure of the subcrustal liquid lava. If they are formed in this way the cavity which is sometimes found be-

neath the crust of the schollendomes is due to a withdrawal of lava after the dome formed.

The walls of many of the cavities shown in Figure 1 of Plate 3 are covered with stalactites coated with a microscopic film of glass. These stalactites formed from lava which dripped from the roofs of the cavities. The surface of the stalactites cooled rapidly forming this film of glass. On the surface of the flow the glass, which may be as much as half an inch thick, is always black, while the glass on the stalactites is invariably green. As the green glass is much thinner the color difference might be due to the difference in the thickness of the two glass films. However, several specimens of the lava were fused in a blast lamp and in every case a black glass formed. As the glass films formed in the blast lamp were as thin as the green glass films, thickness does not seem to be the determining factor. This color difference may be due to a difference in the chemical composition of the two glasses. No difference in the optics of the glasses could be detected, so that if any chemical difference exists it is very small. Such a chemical difference may have been caused by the differences in composition between the gases in the cavities and those in the atmosphere at the time the glasses were formed.

SPATTER CONES

The last 6 miles of the McCartys flow was intensively studied, and in this area only one spatter cone was found. Spatter cones are common on the lava flows of Iceland (Geikie, 1897), Hawaiian Islands (Perret, 1913, p. 152–153; Stearns, 1930, p. 122-123), and elsewhere. The spatter cone on the McCartys flow has an approximately circular base; it is about 4½ feet high and about 10 feet wide. The clots of lava which formed it were all liquid when ejected, and many of the larger ones show evidence of flowage. Most of them are shaped like pancakes. As the cone grew upward its walls converged, forming the bee-hive shape shown in Figure 2 of Plate 3. Apparently the horizontal component of the trajectory of the clots gradually decreased as the cone grew. This decrease is best explained by assuming that the vent out of which the clots were emitted became progressively smaller and deeper as the crust thickened. The walls are thin, and there is a large chamber within the cone. The walls of this chamber are lined with lava stalactites. The smaller ones are commonly dumbbell-shaped due to the solidification of a large drop of lava at the end of the stalactite. The larger stalactites are often covered with small drops of lava, many of which are spherical but some of which are pear-shaped. The stalactites vary from three quarters of an inch to 1½ inches in length and represent the drip from the clots which made the walls of the cone.

CRACKS

Two kinds of cracks are found in the McCartys flow: wedge-shaped cracks which may be as much as 6 feet wide, 13 feet deep, and more than 100 feet long, and much smaller nearly parallel-walled cracks which are in most places too small to map (Fig. 4, F). In many places these parallel-walled cracks have saw-toothed walls because the break took place along columnar joints. The width of the flow measured along

the top of the cross section of Figure 4, G is 190 feet, and the sum of the widths of the cracks shown and not shown in the cross section is about 16 feet.

That the contraction resulting from the change of glassy to crystalline basalt is only partly involved in the formation of these cracks is evident. The volume of any crystal formed in the cooling liquid lava was approximately 10 per cent (Daly, 1933, p. 50) less than the volume of the liquid from which it was derived. However, this crystallization did not necessarily result in a general contraction of the cross section because additional liquid lava may have flowed in as the contraction proceeded. If a crystal mesh was ever developed in which liquid lava was completely surrounded by crystals this residual liquid might crystallize and cause a contraction which could not be offset by the addition of new liquid. However, it is difficult to see how such contraction could produce the type of cracks under discussion, and moreover it is doubtful if such a crystal mesh was ever developed.

The work of Day, Sosman, and Hostetter (Daly, 1933, p. 48) showed that there was about a 3 per cent decrease in volume in the change from crystalline Palisade diabase at 1200°C. to crystalline Palisade diabase at 20°C. The McCartys basalt is comparable to the Palisade diabase and as the McCartys basalt may have crystallized around 800°C. it could have contracted about 2 per cent in cooling from its crystallization temperature to 20°C. As it is 190 feet wide this would account for something less than 4 feet of contraction. Hence the magnitude of these cracks is too great to be explained by thermal contraction alone. Moreover, if they were due to thermal contraction one would expect all of them to have nearly parallel walls.

On the other hand, if the initial transverse section of the flow was approximately as shown in Figure 4, E and if the crust was elevated to its present position either by a constriction which prevented flow downstream or by an increase in the volume of lava coming from upstream, then the crust would be stretched approximately 10 feet by this uplift. This together with the contraction which must have taken place when the lava cooled, from approximately 800°C. to 20°C., can account for the formation of the cracks. The large wedge-shaped cracks were probably produced mainly by the stretching consequent on uplift, while the smaller, nearly parallel-walled cracks are due to thermal contraction. Uplift also explains the present domical cross section of the flow as the transverse section of a basaltic flow usually has steep sides and a relatively flat top.

However, the bottom of the uplifted crust should be stretched practically as much as the top. This should produce not wedge-shaped cracks but parallel-walled cracks which would be as deep as the crust was thick. That the cracks are wedge-shaped may be explained as follows: (1) If the crust acted as a rigid member there would be formed wedge-shaped cracks in the bottom part of the crust opening downward, as well as those in the top part of the crust opening upward. (2) If the bottom of the elevated crust was somewhat viscous, flowage of this viscous material into the bottoms of the cracks might make them wedge-shaped. (3) If the crust was slowly uplifted while it was progressively thickening, the cracks formed should be wedge-shaped as the upper parts of the crust having been crystallized the longest would be stretched more than the bottom part of the crust which would have been liquid during all except the last stages of the uplift.

The wedge-shaped cracks contain banded lava and shark's-tooth projections (Nichols, 1939a, p. 188–192). These phenomena can be explained easily if the cracks were formed by the stretching of the crust, as the liquid lava responsible for the uplift could supply the necessary gases to form the banded lava and the shark's-tooth projections could be formed by movements which resulted from the uplift. On the other hand, it would be difficult to explain them if these cracks were due to thermal contraction. Moreover, the rapid widening of the flow downstream from this cross section may be due to a spreading of the flow consequent upon the same increase in the volume of lava which has been postulated to explain the doming (Pl. 2, fig. 1).

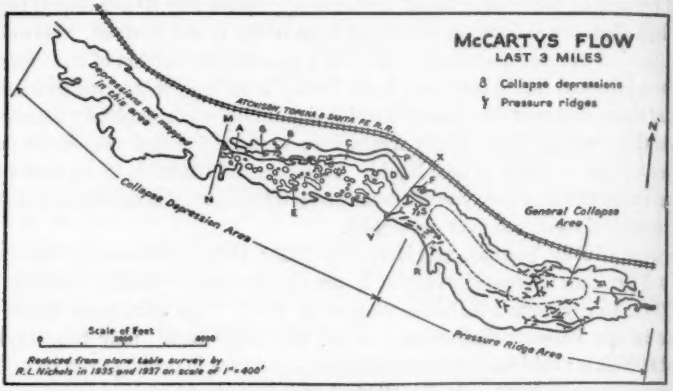
McCARTYS FLOW
LAST 3 MILES
8 Collapse depressions
Y Pressure ridges
ATCHISON, TOPEKA & SANTA FE R.R.
Depressions not mapped in this area
Collapse Depression Area
General Collapse Area
Pressure Ridge Area
Scale of Feet
Reduced from plane table survey by R.L.Nichols in 1935 and 1937 on scale of 1"=400'

FIGURE 8.—*Distribution of collapse depressions and pressure ridges*

Smaller collapse depressions are shown by closed lines. Larger ones are labeled. The length and alignment of pressure ridges are shown by lines drawn along their crests.

COLLAPSE DEPRESSIONS

DESCRIPTION

Collapse depressions are among the most characteristic features of the last 6 miles of the McCartys flow (Fig. 3). They vary greatly in size. The largest depression is nearly a mile long and in places as much as 300 feet wide, while the smallest is only a few feet in diameter. However, most of them are a few hundred feet long, 50 to 100 feet wide, and 15 to 20 feet deep. In shape, too, they vary greatly; some are round and others oblong, while most are quite irregular. The variations in size and shape are shown by Figure 8, a map of the last 3 miles of the flow. In an area about 2 miles long there are approximately 100 collapse depressions, 50 of which have been mapped. This area will be referred to as the collapse depression area (Pl. 4, fig. 1; Pl. 5). The depressions are commonly filled with water or floored with alluvium and vegetation. The long axes invariably parallel the long axis of the flow, and the surface of the flow between the depressions is relatively flat. These depressions are similar to those described by Russell (1902, p. 101; 1903, p. 54; 1905) and others and are formed by the collapse of the roofs of lava tunnels.

PART OF COLLAPSE DEPRESSION AREA OF McCARTYS FLOW

Flow covered with scattered vegetation, mostly sage, 1 to 3 feet high.

FIGURE 1. SMALL POND IN COLLAPSE DEPRESSION
Man seated at end of depression gives scale. Depression bordered by overhanging cliffs and by sloping surfaces.

FIGURE 2. SMALL COLLAPSE DEPRESSION APPROXIMATELY 15 FEET IN DIAMETER

COLLAPSE DEPRESSIONS CONTAINING WATER AND VEGETATION

At the bottom of many of the collapse depressions angular blocks are piled one on the other. That these blocks were derived from the collapsed roof of a tunnel is shown by their vesicular nature. Some of the blocks have a ropy surface such as is characteristic of the surface of the McCartys flow in this area. In by far the greater number of depressions, however, no blocks are visible because the original floor is covered by water, alluvium, or vegetation (Pl. 4, fig. 2; Pl. 6, figs. 1, 2). The height of the water in the collapse depressions is controlled by the height of the water in the Rio San José as is shown by a comparison of Figure 2 of Plate 2 and Plate 5. Figure 2 of Plate 2 shows a portion of the McCartys flow and the flood plain of the Rio San José shortly after a heavy rainstorm. The stream channel is overflowing, the flood plain in the foreground is partly covered with water, and a Y-shaped collapse depression is partially filled, making a small pond within the basalt flow. On the other hand, Plate 5 shows the same area at a time when the discharge of the Rio San José was much lower, and the flood plain and also this Y-shaped collapse depression were dry. The basalt flow is permeable, for even a casual examination of the flow shows it to be filled with ramifying joints and cracks. The depressions near the center of the flow contain the deepest water. The water in the depressions is often quite muddy after a rain. Only a small part of this mud is derived from the drainage basins of the collapse depressions as most of it is the mud carried by the flood waters which have penetrated the basalt flow. It is this mud and silt which is mainly responsible for the burial of the blocks, although wind-borne dust and vegetal remains also accumulate. In general, if the water table of the basalt and near-by alluvium lies below the floor of a collapse depression uncovered blocks are present; if the water table lies above the floor, for even part of the year, the blocks are covered by water, alluvium, or vegetal remains.

Many of these depressions have a characteristic assemblage of water plants, although the general vegetation of the valley is drought-resisting and adapted to about 10 inches of rainfall.

Dr. Richard Goodwin of the Department of Botany at Rochester University has identified the following species from these depressions: (1) *Typha latifolia* L., (2) *Polypogon Monspeliensis* (L.) Desf., (3) *Phragmites Phragmites* (L.) Karst., (4) *Eleocharis montana* (H.B.K.) Roem. and Schult., (5) *Eleocharis rostellata* Torr., (6) *Scirpus occidentalis* (S. Wats.) Chase, (7) *Scirpus Brittonianus* Piper, (8) *Juncus Torreyi* Coville, (9) *Lemna minor* L., (10) *Halperpestes cymbalaria* (Pursh.) Greene, (11) *Hippuris vulgaris* L.

With regard to these plants Dr. Goodwin (personal communication) reports:

> "This association of plants indicates that the habitat is wet throughout the year and that it is rather saline in nature. Similar associations were not observed elsewhere in the valley except in the neighborhood of salty springs."

Each depression has its own characteristic flora depending upon its relation to the water table. The decay of this vegetation generates considerable quantities of marsh gas which is responsible for the disagreeable odor near the depressions. The pools of water also contain small fish, tadpoles, and frogs, and they act as places of refuge for ducks in the migratory seasons.

The depressions in which the blocks derived from the collapsed roof can still be seen may be as much as 28 feet deep. Twelve to 15 feet is a reasonable figure for the depth of those depressions whose blocks are covered (Fig. 9; Pl. 6, fig. 1).

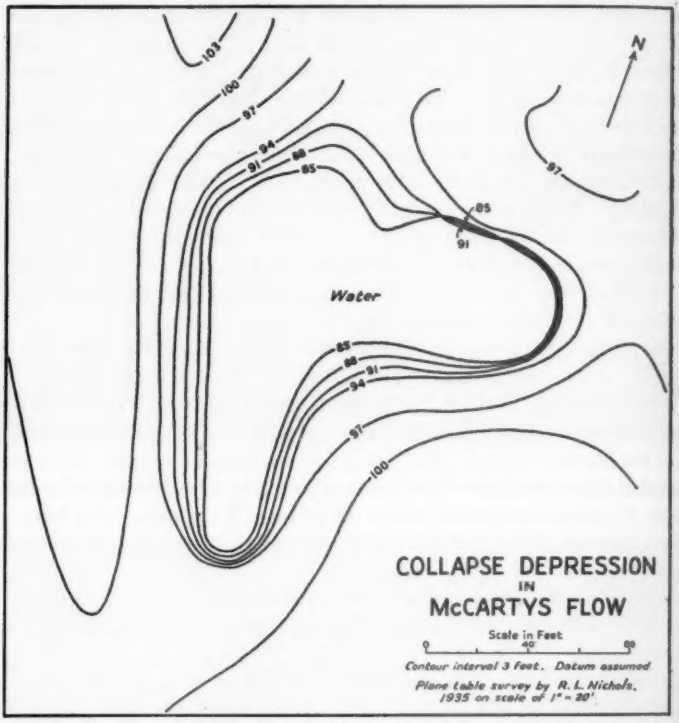

FIGURE 9.—*Topographic map of depression in Figure 1 of Plate 6*

In those collapse depressions where the blocks can still be seen, the depth of the depression gives roughly the height of the lava tunnel, the partial collapse of which produced the depression.

ORIGIN

The distribution of the collapse depressions (Fig. 8) indicates an origin due either to the partial collapse of the roof of one large lava tunnel or of several small distributary tunnels. However, analysis of the probable strength of the roofs of lava tunnels indicates it is unlikely that the roof of a tunnel as large as would be required to span this area could have been maintained even in part. The following calculations lead to this conclusion.

The following assumptions are made. (1) The specific gravity of the upper part of the flow is 2.5; (2) the roofs of the tunnels were, on the average, 8 feet thick; and

(3) the ultimate tensile strength of the not wholly cooled basalt was 1000 pounds per square inch (Swain, 1924), although it may have been lower as the basalt was cracked and jointed. By substituting these somewhat imperfect data in the beam formula (Timoshenko and MacCullough, 1935, p. 111) familiar to engineers:

$$S = \frac{Mc}{I}$$

where S is unit fiber stress, M is the bending moment, I is the moment of inertia of the cross section, and c is ½ the depth of the beam, M can be calculated for the unsupported roof. By assuming that the beam is uniformly loaded and using the value of M thus obtained, the maximum length of beam which could support itself is found to be approximately 100 feet, as calculated from the following formula (Timoshenko and MacCullough, 1935, p. 93):

$$M = \frac{wl^2}{8}$$

where M is the bending moment at the center of the span, w is pounds per linear foot of the beam, and l is the length of the beam.

These calculations assume that the roof was supported only as a simple beam. However, Professor F. N. Weaver of Tufts College says that a roof 8 feet thick, over a tunnel not more than 30 feet high with a curved cross section, would undoubtedly support itself in part by arch action. However, he thinks that a 150-foot span supported in part as a simple beam and in part by arch action would probably be the maximum width which could maintain itself under these conditions.

The length of the span which will support itself is a function of the thickness of the roof and the strength of the crust. However, calculation shows that if the roofs of the tunnels were 24 feet thick, an impossible figure, and if the basalt had a tensile strength of 4000 pounds per square inch, which is much greater than allowable values, the span could not have been as much as 400 feet.

If the collapse depressions resulted from the partial collapse of one large tunnel, this tunnel would have to be as much as 1000 feet wide (Fig. 8). The above calculations show, however, that the roof of such a tunnel would be unable to maintain itself even partially, and that its collapse would result in one wide continuous collapse depression. It would seem, therefore, that the collapse depressions are the result of the partial collapse of many distributary tunnels most of which were probably not much more than 150 feet wide. Collapse occurred where the tunnel was wider or the roof thinner than usual, where the roof was much jointed and cracked, or where the shape of the tunnel did not allow arch action to support the roof. The lava was being supplied to the front of the flow by a system of distributary tunnels as suggested by Figure 6. Moreover, according to Jaggar (1936) this is what is customarily found in the case of the pahoehoe flows of the Hawaiian Islands.

Although the long axes of the collapse depressions, almost without exception, are alined in the direction of the flow, nevertheless, their arrangement does not indicate the number, position, or direction of the tunnels.

The low-lying area—*A*, *B*, *C*, *D* on Figure 8—is covered with alluvium for practically its entire length and apparently resulted from the more or less complete collapse of one continuous lava tunnel. Such a depression or collapsed tunnel has been called a mawai (Jones, 1937, p. 878) by the Hawaiians. This tunnel must have been nearly a mile long, in places as much as 350 feet wide, and oriented in about the same direction as the flow. The peninsular-shaped body marked *P* on Figure 8 might be a lateral tongue which diverged from the flow rather than an uncollapsed portion of the main mass of the flow. If this is correct, the area between the peninsula and the main body of the flow, marked *CD* on Figure 8 is not a collapsed area but is low-lying because it was never covered with lava. However, the outline of the flow at *F* would be difficult to explain, if *P* were a lateral tongue, for it would be necessary to assume that *F* ran back up the valley. Moreover, the absence of such lateral diverging tongues elsewhere on the McCartys flow also makes it doubtful that *P* is a tongue. Also, the shape of the area *CD* indicates that it is a continuation of the undoubted collapsed area *AB*, and this fact strengthens the argument that *CD*, too, is a collapsed area. The ridge labeled *G* in Figure 8, which is found within the collapse area *ABCD*, is either an uncollapsed portion of the roof of the tunnel or a pressure ridge which was formed before the collapse depression.

The main factors which determine whether a roof will support itself or not are its thickness and the width of its arch. It seems unlikely that complete collapse occurred in the *ABCD* collapse depression because the roof of this tunnel was thinner than those of others. On the contrary, the crust of a solidifying flow would normally be thickest near its edge. It seems, therefore, that the complete collapse of the roof of this tunnel must have been due to its great width. However, if for any reason this tunnel was formed before the others it might well have had a thinner crust.

Although the collapse depressions just described are best explained as resulting from the partial collapse of the roofs of many lava tunnels, the depression marked "General Collapse Area" on Figure 8 is apparently due to the collapse of the roof of a large tunnel formed by a general subcrustal drainage. This is indicated by the fact that in these depressed areas the crust is everywhere depressed about the same amount. The surface of these areas as shown by Figure 10 is between 5 and 15 feet below the general surface of the flow. Such general subcrustal drainage is common and examples have been described by Daly (1925, p. 21), Glangeaud (1913), Russell (1903, p. 39–40), and others.

Facing in toward the general collapsed area and also around the depression labeled *ABCD* on Figure 8 are several excellent slump-scarps (Finch, 1933) produced by the collapse.

This marked concentration of collapse depressions, unique along the whole 30 miles of the flow, indicates that the conditions necessary for their formation were not everywhere existent.

Viscous lava, a flat gradient, slow outwelling of the lava at its source, and topographic irregularities on the valley floor would favor the production of many ramifying tubes. These conditions are also those favoring flow by the flow-unit mechanism (Nichols, 1936, p. 629), which is dependent upon the existence of many ramifying tubes. The drainage of these tubes, after the formation of a partially self-supporting crust, would produce the collapse depression area.

Fluid lava, a steeper gradient, and a rapid outpouring of a large quantity of lava would favor the production of one central feeding tube. Such a flow would advance by the single-unit flow mechanism, and as the result of subcrustal drainage a general collapse area would be produced similar to the ones shown in Figure 8. The viscosity and quantity of lava extruded must have been about the same when the lava was advancing over the collapse depression area as when advancing over the general col-

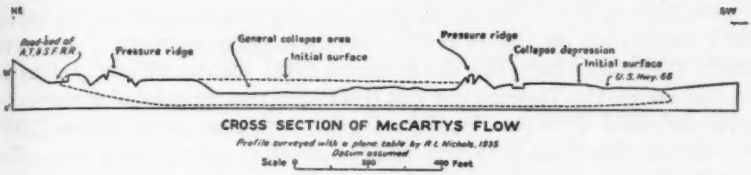

FIGURE 10.—*Cross section of McCartys flow*

lapse area. Therefore, it seems likely that a difference in the gradient of the floor of the valley may have been responsible for the two different kinds of collapses. A steep slope producing the general collapse area and a flatter slope producing the collapse depression area. However, the gradient cannot have been too flat or the tubes, when lava was no longer supplied to them, would not have been drained to form the tunnels.

Drainage of tubes is a prerequisite for collapse. In order that drainage should occur, the flow of the liquid lava from upstream must stop. Such stoppage may be due either to a failure of extrusion of liquid lava at the source or to the obstruction of the channel of flow at some point upstream. Presumably that part of the tube nearest the source or obstruction would be emptied first, and as drainage proceeded the open tunnel would extend itself downstream.

In the triangular area around the kipuka (Fig. 3) the basalt flow probably advanced by means of a number of tongues which coalesced to form this broad triangular lava plain. As depressions are not common above the triangular area, it would seem as if a plugging of the tubes in this area might have been responsible for the drainage of the lava tubes downstream. As the tubes in the triangular area would be smaller than where the flow advanced as one tongue, and as many of them had to make a sharp turn in their progress down the valley, it seems likely that they might have become plugged. No significant collapse occurred between the triangular area and the collapse depression area because the roof was strong enough to support itself.

TIME OF DRAINAGE OF TUBES

The roof of only one lava tunnel (Fig. 8) was more or less completely collapsed indicating that the crust had considerable thickness at the time of the withdrawal of the lava from the lava tubes. Had the crust been only a few inches thick, the roofs over the drained tubes would have everywhere collapsed, and elongated depressions similar to the one marked ABCD on Figure 8 would have been produced. The size and quantity of the blocks found in the bottoms of some of the depressions also indi-

cate a considerable thickness of the crust at the time of the withdrawal. That thick crusts had been formed by the time of withdrawal is further indicated by a cave found in the flow, the roof of which is 8 feet thick. This figure gives only a minimum thickness for the crust at the time of the withdrawal, for the cave has undoubtedly enlarged itself upward by the falling of blocks from the under side of its roof.

If it is assumed that the crust was 8 feet thick at the time of the withdrawal of the lava, that the necessary stiffness had been achieved at 900°C. (Daly, 1933, p. 66), that the flow was 32 feet thick at this point, that the temperature of the top and bottom of the flow was maintained at 0°C., that the initial temperature of the lava was 1200°C. (Verhoogen, 1939; Day and Shepherd, 1913), that the diffusivity of the lava was that assumed by Thomson and Tait (1883), and that the point of slowest cooling was at the middle of the flow; then by using the method and curves of Lane (1898) it is found that approximately 16 days would be necessary to form such a crust. However, if it is assumed that the point of slowest cooling was about a third of the way from the bottom of the flow to the top, as this is usually the true condition according to Prof. A. C. Lane (personal communication), and that 2 feet of the roof had fallen to 900°C. and was therefore practically crystallized, and that the other 6 feet was a stiff glass that could maintain itself at the time of the withdrawal of the lava, then Lane's curves (1898) show that such a roof would have been formed in approximately 17 hours. These calculations assume a closed cooling system during the formation of the crust. However, this was not the case as liquid lava was undoubtedly flowing beneath the crust as it was slowly thickening. This would increase the time necessary to form the crust by a small amount, and because of this fact and the necessity of making so many assumptions this figure should be taken only as indicating the order of magnitude of the time involved.

These calculations indicate that approximately 17 hours elapsed between the arrival of the molten lava at the point where the cave is found and the time when that part of the tube in which the cave is found was drained.

The drainage of this tube caused further advance of the flow down stream and as the end of the flow is 1½ miles from the cave it appears that more than 17 hours were required for the flow to advance this distance. The flow was therefore, moving slowly. Some of the Hawaiian flows have been known to advance so rapidly that no real crust was formed until the front had progressed several miles past the point of crust formation.

TIME OF FORMATION OF THE DEPRESSIONS

Collapse of the roofs of lava tunnels may occur long after the drainage of the tubes and complete solidification of the flow (Russell, 1902, p. 101; Williams, 1932).

Some of the depressions on the McCartys flow no doubt have been formed quite recently; indeed, collapse depressions could be produced at the present time if heavy loads were placed on the roofs of some of the existing tunnels. On the other hand, in some localities, collapse occurred while liquid lava was still present and when the under side of the crust was still soft and plastic.

Thus a cavelike collapse depression was found in the McCartys flow which has a

vertical opening about 5 feet in diameter and 5 feet long. The lower 2½ feet of the opening is grooved and striated. These grooves and striations could have been formed only when the lava was soft and plastic. Soon after the lava tunnel was formed, a portion of its roof collapsed. The upper part of the crust was solid, whereas the lower part was still soft and plastic. When the upper solid part of the crust in falling scraped across the lower plastic part of the crust, grooves and striations were produced (Nichols, 1938, p. 607–608). This collapse took place, therefore, after the drainage of the tube and while the lower part of the crust was still soft and plastic, or, according to the calculations above, about 17 hours after the lava first arrived at this point.

The collapse depression labeled *E* on Figure 8 is of average width and length, but the roof, instead of having collapsed 15, 20, or 25 feet, as is the case with other depressions, has fallen only between 5 and 10 feet. That liquid lava was still present when collapse first occurred is proved by the following: (1) Within and around the margins of this collapse depression are an unusual number of squeeze-ups (Nichols, 1939b) which were extruded from the cracks in the crust produced by the collapse; (2) the squeeze-ups ran toward the depression, showing that they were extruded after the collapse began; and (3) the squeeze-up lava in places has partly buried fragments of the crust broken by collapse.

On the other hand, there appears to have been a small amount of additional collapse after the extrusion of the squeeze-ups, which is indicated by their joint systems and by the fact that these squeeze-ups have been bent downward after their formation. Apparently the collapse occurred while the lava in the tube was slowly being drained away.

The shallow depth of this depression, 5 to 10 feet, may be due to the fact that this tube was not so large as the other tubes, or that drainage was not complete, or that the extrusion of the secondary lava, together with a thickening of the crust by the addition of lava from below, within a brief interval after the initial collapse may have so stiffened the crust that significant additional collapse was impossible.

VOLUME OF THE DRAINED LAVA

Calculation shows that a drainage of lava equal in volume to about 1900 feet of the lower end of the flow was responsible for the lava tunnels in the "Collapse Depression Area" shown on Figure 8.

COLLAPSE DEPRESSIONS AND KIPUKAS IN CROSS SECTION

The cliff which faces in toward a collapse depression is often overhanging; however, the edge of the lava which surrounds a kipuka does not overhang. In cross section a collapse depression consists of a floor of unbroken lava, the bottom crust of the flow, above which there is a chaotic layer of blocks of lava derived from the collapsed roof of the tunnel. The greater part of the area of a kipuka, on the other hand, contains no lava from the flow which formed it, except possibly a few blocks which rolled off the edge of the flow a short distance into the kipuka. The cross section of a kipuka or collapse depression if buried by a later flow might be outlined by a layer of vesicular lava at the bottom of the later flow.

PRESSURE RIDGES

DESCRIPTION

Pressure ridges are common features on many lava flows. They have been described and photographed by Russell (1902, p. 95–97; 1903, p. 54), Stearns and Clark (1930, Pl. 14A), and others. Some of them are similar to the pressure ridges found on sea-ice (Wright and Priestley, 1922, p. 341–354; Leffingwell, 1919; Stefansson, 1910). Such ridges are the most spectacular feature on the McCartys flow. Those on the last mile of the flow were studied in detail (Fig. 3). The shortest is only 130 feet long, while the longest is more than 1200 feet. They vary in height from 10 to 25 feet and are as much as 100 feet wide. However, the pressure ridges of this area are relatively small, as some of those near the source of the McCartys flow are more than 40 feet high. In transverse section the sides of the pressure ridges are steep, suggesting the gable of a house or the cross section of a broken anticline. In no place was overthrusting of one side of a ridge onto the other observed. These pressure ridges have medial cracks running along the crest of the ridge, shown by Figure 1 of Plate 7, a transverse view of one of the 50 pressure ridges on the last mile of the McCartys flow. These medial cracks are usually less than 15 feet wide. The brittle lava on most of the pressure ridges is also broken up into many secondary cracks (Pl. 7, fig. 2). The lava surrounding this pressure ridge is buried by a thin veneer of alluvium. This ridge, although spectacular, is not so representative of the pressure ridges of the McCartys flow as those which are longer in proportion to their width. Figure 11 shows that its sides are, in places, as steep as 45°; most of the ridges, however, are flatter. The position of this ridge on the last mile of the McCartys flow is indicated by *R* on Figure 8. Although most of the pressure ridges consist of a single ridge somewhat resembling an esker, many have secondary ridges branching out at various angles (Fig. 8).

The pressure ridges may be divided into two types (Fig. 8): longitudinal pressure ridges with axes more or less parallel to the flow, and transverse pressure ridges with axes more or less transverse to it. The longitudinal pressure ridges are more numerous and longer and in general are close to the margin of the flow. Although the shorter ridges are usually straight the longer ones invariably are gently curved with a sinuous ground plan. In a few places several parallel longitudinal ridges occur, and occasionally they are arranged en echelon (Fig. 8).

Many of the medial cracks of the pressure ridges have a surficial layer or veneer on their walls (Pl. 7, fig. 1). The veneer consists of bands of dark and less-dark material that extends in from the faces of the cracks not more than one eighth of an inch and which are more or less parallel to the surface of the flow. The bands are 1 to 4 inches wide and scores of feet long. The bands consist of glassy lava alternating with bands of duller, less-glassy lava, the whole forming a veneer on the crystalline basalt that normally forms the walls of the cracks. Usually the glass is black but may also be red, purple, bluish black, or green. This banding described by Nichols (1939a, p. 188–191) is undoubtedly due to refusion of the walls by the oxidation of combustible gases emitted from the base of the cracks, the glassy bands being more thoroughly re-fused than the duller bands.

FIGURE 1. TRANSVERSE VIEW OF PRESSURE RIDGE ON MCCARTYS FLOW
Lava on the walls of the medial crack is banded. Bands can best be seen above and to the right of the rod. Bands should not be confused with lava ropes to right and left of rod.

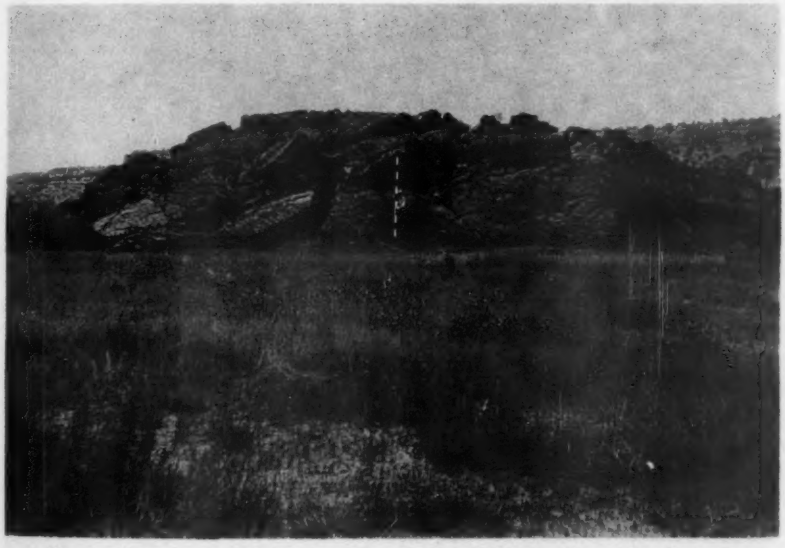

FIGURE 2. LONGITUDINAL VIEW OF THE PRESSURE RIDGE SHOWN IN FIGURE 1 OF PLATE 7

TRANSVERSE AND LONGITUDINAL VIEW OF PRESSURE RIDGE

FIGURE 1. ROPY BILLOWY LAVA WHICH FLOWED OUT OF MEDIAL CRACK OF PRESSURE RIDGE

FIGURE 2. SECONDARY LAVA IN MEDIAL CRACK OF A PRESSURE RIDGE

SECONDARY LAVA IN MEDIAL CRACKS OF PRESSURE RIDGES

TIME OF FORMATION

The sides of these ridges are often covered with ropy lava whose orientation and arcuate form bear no relation to the pressure-ridge topography, proving that the

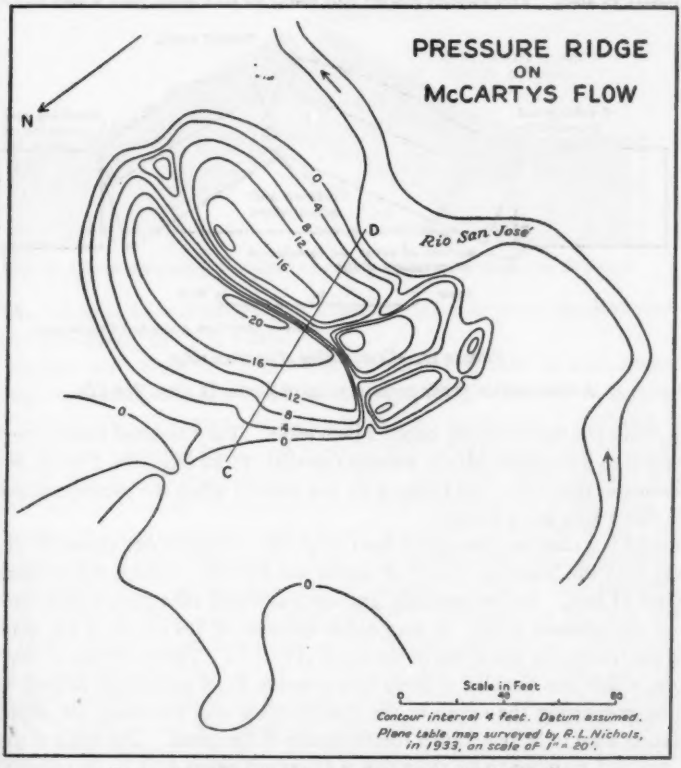

FIGURE 11.—*Topographic map of pressure ridge in Plate 7*

ridges formed after the crust solidified. Most of the medial cracks in the pressure ridges are empty, though some are filled in part with secondary lava (Pl. 8). The bulbous shape of the lava extruded from the medial crack of the pressure ridge indicates that it was quite viscous when extruded (Pl. 8, fig. 1). The viscosity was no doubt due to rapid cooling after extrusion because of the small quantity. That the pressure ridges were more or less completely formed at the time of the extrusion of the secondary lava is indicated by its lack of cracks and joints as compared to the lava of the pressure ridges. The absence of secondary lava from many of the medial cracks may be due to one of the following: (1) Liquid lava was no longer present when these cracks were formed; (2) the cracks were not continuous from the top of the crust to the bottom; (3) the lava was too viscous to move up into the cracks; (4) the top of

the liquid lava did not reach the bottom of the medial cracks; (5) the hydrostatic head of the lava was not sufficient to move it up into the cracks.

The medial cracks of many pressure ridges also contain wedges of lava, the thin edges of which point upward. These wedges are vertically grooved and fluted and vary greatly in size. The largest are 30 feet long, 10 feet high, and 4 feet thick at the

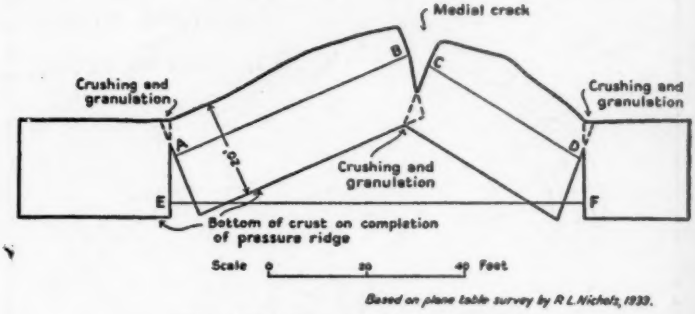

FIGURE 12.—*Cross section of pressure ridge*

A cross section of the pressure ridge of Figure 11 along line *CD*.

bottom, while the top is usually about 1 inch wide. They resulted from the extrusion of viscous lava into progressively widening medial cracks (Nichols, 1938, p. 609–613). They indicate, therefore, that liquid lava was present when the pressure ridges which contain them were being formed.

Figure 12 is a cross section along line *CD* of the pressure ridge shown in Figure 11. It shows that the length of the crust which was buckled to form this pressure ridge was about 87 feet. In the buckling, granulation would take place in the three major joints of the pressure ridge. A reasonable estimate of the depth of the granulation zone is one fourth the thickness of the crust (Fig. 12). The thickness of that part of the crust which was buckled to form this pressure ridge was about 20 feet, as determined by measuring the depth of the medial crack and assuming the depth of the granulation zone to be one fourth the thickness of the crust. The ratio of the length of the buckled member to its thickness is therefore about 4 to 1. A column of concrete, granite, basalt, or any other similar material with a ratio of 4 to 1 will not fail by buckling but rather by shearing or by crushing. The minimum ratio of length to thickness which will yield by buckling is usually said to be about 10 to 1 (Dunham, 1939) and even with this ratio, shearing or crushing rather than buckling is liable to take place. However, if a compressional member has a ratio of 15 or more to 1 it will usually fail by buckling. This ratio applies to perfectly straight, homogeneous, vertical columns. That the lava crust is not perfectly straight should not seriously affect the ratio as the deviation from perfect straightness is never great. Nor should the fact that the crust is cracked and jointed affect it, as these cracks and joints would be closed as soon as the crust was placed under compression. As the crust during the period of solidification was progressively weaker with increasing distance from the surface of the flow, the tendency was for the crust to buckle upward. However, as the

lava crust is a horizontal column, its weight tends to prevent upward buckling. This tends to increase the ratio necessary to insure buckling, but this factor is probably offset by the fact that the nonhomogeneity of the crust tends to lower it. As the lava crust was cracked, jointed, and soft at the bottom, it was undoubtedly considerably

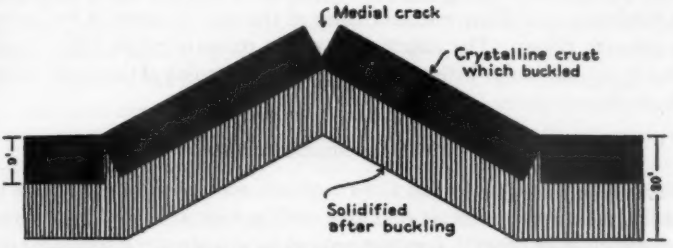

FIGURE 13.—*Medial crack produced by buckling of thin crust which later thickened*

weaker than the solid homogeneous basalt used for building purposes; however, this should in no way affect the 10 to 1 ratio.

If we assume that when the crust first started to buckle it had about the same strength as cold basalt, it is found by applying the 10 to 1 ratio to the pressure ridge of Figure 11 that the crust could not have been more than 8 or 9 feet thick. This means that the crust has thickened progressively from about 9 feet to its present thickness of 20 feet since it first started to buckle. Moreover the crust may be more than 20 feet thick because of thickening following the completion of buckling. The grooved squeeze-ups and secondary lava found in the medial cracks of the pressure ridges indicate that the crust was in contact with liquid lava making possible progressive thickening during the period of pressure-ridge formation. That the crust had considerable thickness when buckling first started is indicated by the fact that no secondary lava broke out from the sides of any of the pressure ridges. Moreover, it is doubtful if it could have been less than 6 feet thick at the time of buckling, as a column much thinner than this would break in spans shorter than 87 feet. The depth of the medial crack indicates that the pressure ridge was slowly raised while the crust thickened from 9 to 20 feet. If the pressure ridge had been raised to its maximum height while the crust was 9 feet thick, and if the crust was then progressively thickened to 20 feet, the medial crack would not have had its present depth but would have been about 6 feet deep and similar to the one shown in Figure 13.

By making assumptions similar to those above and using Lane's curve (1898), it is found that several days were required for the crust to thicken from 9 to 20 feet. This means that this pressure ridge was several days in forming, and, as the other pressure ridges are similar in essential characteristics, very likely all the pressure ridges on the McCartys flow were several days in forming.

CRUSTAL SHORTENING INVOLVED IN FORMATION OF PRESSURE RIDGES

The difference in length between the sum of lines *AB* and *CD* and line *EF* in Figure 12 gives a figure for the crustal shortening involved in the formation of this pressure

ridge. This difference indicates a crustal shortening of 3.25 feet. However, this pressure ridge is not only shorter and higher than most of the others, but it contains a wider medial crack, so that 3.25 feet is perhaps too high a figure for the crustal shortening involved in the formation of the average pressure ridge. If 1.5 feet is the probable shortening Figure 8 shows that 2 to 4 feet would be a good figure for the maximum total transverse shortening of the crust involved in the formation of all the pressure ridges. The distribution of the pressure ridges (Fig. 8) and their variation in size and shape indicate that the total shortening of the crust varied somewhat from place to place.

ORIGIN OF PRESSURE RIDGES

As these pressure ridges are not to be confused with schollendomes (lava blisters) it is hardly necessary to point out that they could not have resulted from the accumulation of gas pressure beneath the crust or from local hydrostatic pressure exerted by fluid lava beneath the crust.

Bradley (1873, p. 204–205) in discussing the low mounds (probably pressure ridges) on the Snake River Lavas, writes:

> "Whatever the source (referring to the lava of the Snake River Plains), the material had evidently become quite viscid; for, at some points, where it ran over small inequalities of the surface beneath, it now stands in low mounds, which would not have been the case if it had been very fluid. That these mounds were not all formed by an undermining and sinking of the surrounding mass, to which some of them have very properly been referred, is proved by the tapering shape of the closely-fitting blocks which form the arch. But there is still room for study on all these points."

That the pressure ridges of the McCartys flow are not the result of the mechanism suggested by Bradley is indicated by: (1) The theory demands an improbable pre-McCartys topography; (2) if such topographic irregularities existed, they would undoubtedly have been buried by the lava without forming surface features on the flow.

That these ridges were not formed by a withdrawal of lava and a collapse of the surface except where supported by pre-McCartys ridges is proved by the following: (1) This theory demands an improbable pre-McCartys topography; (2) the surface of the flow on one or both sides of many of the ridges gives no indication that its present position resulted from collapse; (3) in no place where collapse has occurred on one or both sides of a pressure ridge can the collapse be considered to have formed the ridge; and (4) the fact that the medial cracks are not open at the bottom and that there are no open cracks marginal to the ridges is better explained by compressional forces acting on the surface of the flow than by tension resulting from differential collapse. However, this mechanism has been used to explain certain so-called pressure ridges in ice observed by Wright and Priestly (1922, p.356) in the Antarctic.

In describing the lava of Red Mountain in Lassen Volcanic National Park, Williams (1932) writes:

> "Here may be found a remarkable group of collapsed lava tubes, now represented by a series of north-south ridges separated by gullies from 10 to 50 yards in width, partly infilled with angular blocks through the cave-in of tubes .... Where the lava tubes are closely spaced, the intervening ridges are roughly V-shaped in section, due to the arching of the lava over the tubes."

That the pressure ridges of the McCartys flow were not formed in this way is suggested by the fact that these ridges are not associated with collapse features

similar to those described by Williams, and by the fact that ridges formed in this way would not have medial cracks such as are found in the pressure ridges on the McCartys flow.

Russell (1902, p. 95–97), in discussing the pressure ridges of the Snake River Plains, suggests that they were produced by lateral pressure acting on the surface portion of the lava flows, arising from the viscous drag of slowly moving subcrustal lava. In this respect, states Russell, they are similar to the ropy texture so common on the surface of pahoehoe flows which, as is well known, is due to the viscous drag of slowly moving subcrustal lava on a plastic crust.

The author believes that the pressure ridges on the McCartys flow are the result of lateral pressure acting on the crust. Lateral pressure is not the result of viscous subcrustal drag on the crust for, if a viscous subcrustal drag were the force that formed the pressure ridges, one would expect them to be oriented in general transverse to the direction of the flow. However, the pressure ridges on the McCartys flow are in general oriented with their long axes parallel to the flow. The size and shape of the pressure ridges near the end of the flow are similar to those farther upstream. As the ridges upstream were formed when the crust was several feet thick, a similar thickness is indicated for the crust from which the pressure ridges near the end of the flow were formed. However, as the flow probably reached its terminal position long before this part of the crust could have been thickened so much, it would seem that viscous drag could not have been operative when these ridges were formed. It seems, therefore, that some other mechanism must be considered as responsible for these pressure ridges.

Because the crust of the flow was a continuous carapace of solid lava which surrounded the liquid thread, the top crust was everywhere tied to the margins and bottom of the flow. Before viscous drag could buckle the crust, the top of the crust had to be sheared free from the margins and this block had to be dragged free from the top crust of the upstream part of the flow. Then viscous drag would have to move this block downstream against the unsheared uppercrust of the downstream part of the flow. The downstream part of the dragged member might then buckle itself against this abutment if the viscous drag were strong enough (Fig. 14). Moreover, in the early stages of the process the viscous drag would in all probability have to shear and buckle simultaneously.

The following calculations show that the viscous drag would probably not be strong enough to shear the top crust from the margins of the flow. If we assume: (1) that the liquid thread running within the carapace was 25 feet thick (only the upper 12.5 feet of the liquid thread dragged at the bottom of the top crust as the lower 12.5 feet dragged on the bottom crust); (2) that the specific gravity of the liquid lava was 2.5; (3) that the crust was 8 feet thick at the time of shear (that the crust was sheared when it was much thinner than 8 feet and then remained free while the crust thickened to 8 feet seems unlikely, as the injection of liquid lava from below into the shear planes and tension cracks would in all probability refasten the sheared member back into the carapace); (4) that the gradient of the flow was about 30 feet per mile; (5) that the viscous drag was equal to that component of the weight of the lava which acted downstream;—then calculation shows that the viscous drag on the bottom of

an area 1000 feet across the flow and 1 foot up and down the flow was 4.8 tons. To get the total force, we must add to this the component of the weight of the dragged unit of crust which acted downstream. If the specific gravity of the solid crust was 2.5, then this force is 3.10 tons. The total shearing force is therefore approximately

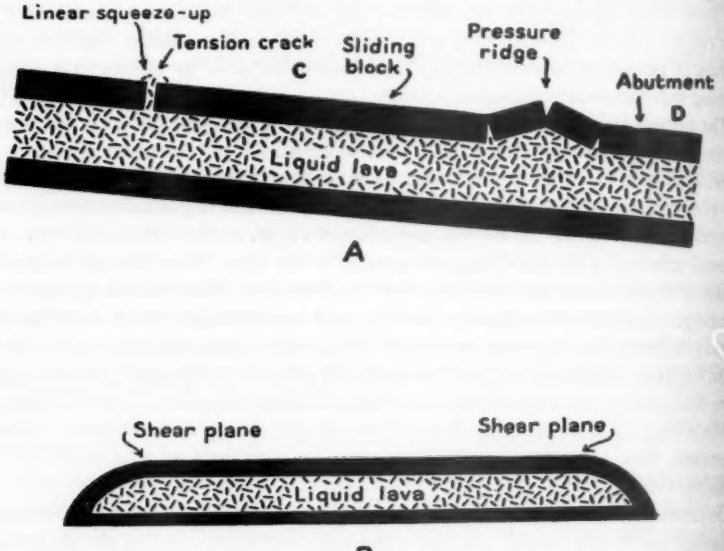

FIGURE 14.—*Formation of pressure ridges by viscous drag*

A is a longitudinal section of the flow showing the formation of pressure ridges by viscous drag. That part of the crust labeled *D* was held rigid and acted as an abutment against which *C* was dragged. The downstream portion of *C* was buckled against this abutment. B is a transverse section of the flow showing the location of the shear planes.

7.9 tons. As the dimensions of the planes at the two margins of the flow which must be sheared by the unit area under consideration are 8 feet by 1 foot, the total area to be sheared is 16 square feet. The force acting on each square foot is therefore approximately .5 ton. The shearing strength of good quality basalt used for building purposes is probably about 150 tons per square foot (Merriman and Wiggin, 1930). Although the crust was undoubtedly jointed, cracked, and hot on the bottom, it seems very unlikely that its shearing strength was as low as .5 ton per square foot. As the viscous drag would always be lower than that calculated, and as no allowance was made for the fact that the crust had to be pulled apart and probably buckled while it was being sheared, these calculations show that there is a very strong probability that viscous drag cannot shear the crust and therefore could not have formed the pressure ridges.

If, however, we assume that a block of the top crust could be sheared from the margins and dragged free from the crust upstream, then the following calculations show that buckling could not have taken place.

If we assume that the block which was dragged and which buckled on its downstream end was 12,000 feet long, 1000 feet wide, and 8 feet thick, then the total force acting at the end of the block would be 94,800 tons. This force would be applied on a surface 1000 feet wide and 8 feet deep so that the pressure would be approximately 11.8 tons per square foot. As the end of the flow is not straight (Fig. 8) the component of the drag which is effective in buckling the downstream end of the block must be considerably less than the total drag. The compressive strength of good quality basalt used for building purposes is approximately 1400 tons per square foot (Merriman and Wiggin, 1930). In view of this, it seems unlikely that the compressive strength of the crust was as low as 11.8 tons per square foot, so that these calculations indicate that viscous drag could not buckle the crust and therefore the pressure ridges could not be formed in this way.

DOMICAL THEORY FOR ORIGIN OF PRESSURE RIDGES

While lava was supplied from upstream to the lava tubes in the collapse depression area, the front of the flow advanced down the valley. A time finally came, however, when lava was no longer supplied to these tubes. At this time the flow, in the pressure ridge area, undoubtedly had steep marginal sides and a flat top. This is suggested by the fact that this type of cross section is found along practically the entire extent of the flow. Following this, the lava in the tubes slowly drained away. Possibly most of this lava made room for itself by doming up that part of the flow which is now occupied by the general collapse area (Fig. 8). The marginal portions of this part of the flow were not domed as the crust was too rigid. During the formation of the dome, tension cracks tended to form from which liquid lava might have been extruded. This secondary lava was not found, however, as that part of the flow where it would be expected is in large part covered with alluvium.

While the dome was forming the crust tended to slide down the sides of the dome. This sliding force was greatest when the dome was highest, and the tendency to buckle because of this sliding was greatest at the bottom of the dome. The sliding of a lava crust on viscous subcrustal lava has been described by Daly (1925, p. 18–20), and it results in the buckling of the crust downstream and in the formation of tension cracks upstream. The buckling might produce pressure ridges; however, no tension cracks or trenches formed as the result of the pulling apart of the crust at the top of the dome were found. It will be shown below that the force resulting from the tendency to slide on a dome of any reasonable height is not strong enough to buckle a crust of the necessary thickness.

When the crust was between 6 and 9 feet thick the lava in the dome started to drain downstream. Continued drainage left more and more of the dome unsupported. Finally the crust could no longer support itself and collapse took place. However, the crust was too wide to accommodate itself to its new position and it buckled near the margin of the dome. Continued drainage caused continued collapse and the ridges formed by the buckling slowly grew. Whether the dome fell gradually or with a series of jerks would depend upon the way the lava was drained and upon the strength of the crust. As the pressure ridges were days in forming the lava must have drained very slowly from under the dome. The variations in the size and shape

of the pressure ridges resulted because the crust was not homogeneous and the dome varied in dimensions from place to place. As the highest part of the dome would produce on collapse the greatest shortening, the pressure ridges marginal to this area should be more numerous or higher than elsewhere. The collapse of such a dome would result in compressional forces perpendicular to its margin; the pressure ridges formed would therefore in general be parallel to the margin of the dome. That the ridges should be close to the margin of the dome is due to the following: (1) The compression resulting from collapse is at a maximum near the margin; (2) the compression resulting from sliding is also at a maximum near the margin; and (3) the hydrostatic pressure of the liquid lava, which is directed vertically and therefore aids buckling, is also at a maximum in the area marginal to the dome.

Finally, the dome became flat and the pressure ridges reached their maximum size. Further drainage produced the general collapse area and was responsible for the fact that a few of the pressure ridges have in part fallen into collapse depressions. As the crust was in tension rather than in compression at this time, the medial cracks of some of the pressure ridges may have slowly opened, so that the liquid lava which was still present below could be intruded, forming the grooved squeeze-ups and the fluid secondary lava. The thickness of the grooved squeeze-ups at their bases indicates the amount of separation of the sides of the pressure ridges which took place during this period of relaxation.

In many places the grooved squeeze-ups and the secondary lava filling the medial cracks are higher than either the present general collapse area or the initial position of the crust. However, the collapse from the horizontal position to the present position probably could have squeezed up the liquid lava to this height. Moreover, the squeeze-ups and the secondary lava may have been formed when the dome was high. Although in general the medial cracks would be closed at this time, yet parts of some of them might have opened, especially if there was any lateral movement of one side of the pressure ridges past the other and if the medial cracks were not straight. The height of the squeeze-ups and secondary lava can easily be explained by this mechanism as the lava in the dome would be higher than the pressure ridges. The medial cracks are not straight and it seems likely that some lateral movement of the sides of the pressure ridges did occur during growth.

Figure 15 is a series of sketches illustrating the formation of these pressure ridges. *A* is the initial cross section of the flow. *B* is the cross section after the dome was formed. The formation of embryonic pressure ridges marginal to the dome due to partial collapse is shown in *C*. *D* represents a more advanced stage of collapse, while *E* is the final stage—the general collapse area, the pressure ridges, and the grooved squeeze-ups have been formed, and the flow is completely solidified.

The pressure ridges which are marginal to the flow and lined up longitudinally with it were formed by compressional forces acting transverse to the flow, which resulted from the collapse of the dome. Compressional forces acting longitudinal with the flow were also produced by the collapse of the dome. These forces would form pressure ridges transverse to the flow. Those ridges which are transverse to the main direction of the flow and near the area labeled *S* on Figure 8 were formed by compressional forces resulting from the collapse of that part of the dome which faced

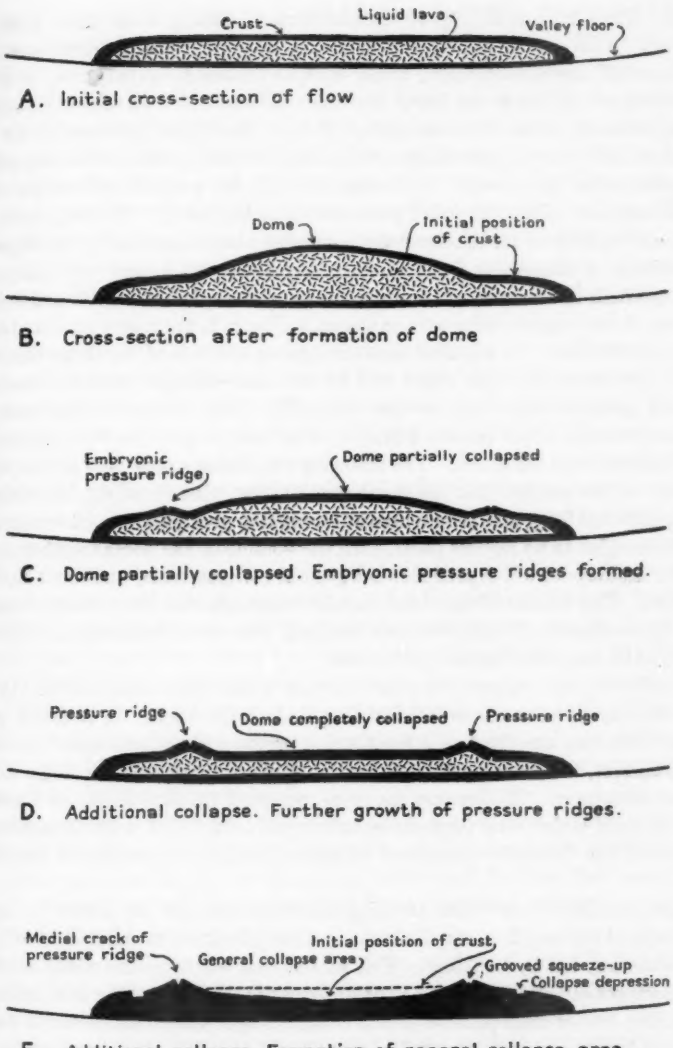

FIGURE 15.—*Formation of pressure ridges by collapse of dome*

upstream. The alignment of several of these ridges is difficult to explain by the collapse of the dome; however, the cracks, joints, and general heterogeneity of the crust are probably responsible. The ridges labeled *L* on Figure 8 which are nearest the end of the flow resulted from the collapse of that part of the dome which faced down-

stream. Those labeled *K* (Fig. 8) resulted from a buckling of the crust of the top of the dome. This may have been due to the fact that the crust was weaker here than elsewhere and therefore buckled when collapse occurred; or they may have been formed because the dome was lower here than elsewhere and was therefore subjected to compressional forces directed against it from the higher portions of the dome. The formation of the pressure ridges at *K* on the downstream side of a collapsed dome is an alternative hypothesis. Following the collapse, a smaller dome was formed downstream, the collapse of which produced those labeled *L*. The drainage of these two domes would produce the two general collapse areas separated by the ridges at *K*. The drainage of the smaller dome advanced the flow to its present terminal position. This hypothesis is, however, not favored, as the other explains the facts and is simpler. In a part of the pressure ridge area as shown in Figure 8, the ridges are found on only one side of the flow. If the crust was stronger on one side of the dome than on the other in this area—and this might well be the case—collapse probably would have produced pressure ridges only on the weak side. The volume of lava downstream from the pressure ridges labeled *L* should be at least as great as the volume of lava which drained from the dome. The following calculations show that this is probably the case. If we assume that the dome was 800 feet wide (it might have been narrower), 4000 feet long, and 27 feet high at its highest point, and that it rose gradually from its margins to its highest point, then the volume of the dome together with the lava, the drainage of which produced the general collapse areas, was about 30,000,000 cubic feet. The volume of the visible lava downstream from the pressure ridges at *L* is also approximately 30,000,000 cubic feet, and there is undoubtedly a considerable volume which has been buried by alluvium.

The following facts suggest that the existence of this dome was possible: (1) Schollendomes may be as much as 100 feet long and 15 feet high. Daly (1933, p. 155) believes that they are due to the hydrostatic pressure of the subcrustal liquid lava. If he is correct, the dome postulated for the origin of these pressure ridges is a very large schollendome. (2) Because the cross section of the flow shown in Figure 4, F is the result of hydrostatic pressure by subcrustal lava, belief is strengthened in the existence of the dome the collapse of which is thought to have formed the pressure ridges.

Several parallel ridges which are longitudinal to the flow are shown in Figure 8. If the ratio of the length to the thickness in a compressional member is 10 or 15 to 1, it will generally fail by buckling. This means that wherever the width of the area under compression was more than 30 or 40 times the thickness of the crust at the time of buckling, two or more parallel pressure ridges might have been formed if the compressional forces were strong enough to cause buckling. Whether these parallel ridges were formed contemporaneously or whether those nearest to the center or nearest to the edge were formed first cannot be analyzed, as the strength of the crust and the way in which the dome collapsed are not known.

Many of the pressure ridges are sinuous in ground plan (Fig. 8). The lines of equal strength in the flow would at no time be straight because the margin of the flow is not straight, the thickness of the flow is not uniform, and the joints and cracks in the crust are irregularly distributed. Moreover, the outward thrust of the unsupported dome

would vary from place to place because the height and width of the dome would vary and because the thickness of the crust in the dome would also vary. The combination of these factors would result in pressure ridges with a serpentine ground plan.

An analysis of the distribution of the pressure ridges indicates that the dome would have to be about 800 feet wide and nearly a mile long in order to account for them. The approximate outline of the dome is shown by dotted lines in Figure 8. The pressure ridges involved a crustal shortening of 2–4 feet. If the dome were 27 feet high with a base 800 feet wide, the distance along the dome would be about 802 feet. The collapse of this dome would produce then about 2 feet of shortening. If the dome were 38 feet high, 4 feet of shortening would result on collapse. A dome, therefore, between 27 and 38 feet high could account for the necessary shortening on collapse.

The pressure at the base of the dome due to the sliding action of the crust over the liquid lava is calculated by resolving the vertical weight of the crust into components —one perpendicular to the surface of the liquid lava and one parallel to it. The force parallel to it is the one tending to buckle the crust at the base of the dome. If we assume a frictionless base under the crust, a dome 27 feet high with a base 800 feet long, a specific gravity for the solid basalt of 2.5 (low because of vesiculation), and that the crust of the dome was about as thick as that where the pressure ridges were formed, calculation shows that this force would be about 2.1 tons per square foot. The crushing strength of basalt is about 1400 tons per square foot. As the crust of the flow is jointed and cracked, a good figure for its crushing strength is probably about 350 tons per square foot. Although the force necessary to buckle the crust was much less than the crushing strength of the material (Timoshenko and MacCullough, 1935, p. 259–271, p. 278–282), it could hardly have been as low as 2.1 tons per square foot. These calculations show, then, that the sliding force on a dome 27 feet high is insufficient to buckle the crust and form these pressure ridges. If the dome were 38 feet high, the sliding force would still be too small to cause buckling, as it would be only about 3.0 tons per square foot.

The dome is supported only by the sides of the flow after the drainage of the liquid lava from beneath it. An arch is therefore formed, the abutments of which are the sides of the flow. As the crust is thickest at the sides of the flow, due to more rapid cooling, it seems reasonable to suppose that these abutments were strong enough to carry the horizontal thrust of the arch.

If it is assumed that the specific gravity of the solid crust was 2.5, that the crust was 8 feet thick throughout the dome, and also at the point of buckling at the base of the dome, that the dome was 800 feet wide and 27 feet high, then the average pressure at the base of the dome as computed by the method of analysis of a three-hinged arch, assuming hinges at the crown and at the abutments, is found to be about 232 tons per square foot. As the thrust at the base of the dome probably does not act at the center of the crust, the pressure at the top of the crust is likely to be at least twice the average pressure or 464 tons per square foot. As the crushing strength of the crust was probably about 350 tons per square foot and as the force necessary to cause buckling was much less than the crushing strength, it is apparent that the dome when unsupported would collapse and buckle the crust. If the dome

were 38 feet high, the average pressure at the base would be 165 tons per square foot. Such a dome might well collapse, but there is less certainty of collapse than in a dome 27 feet high. While this method of analysis is not exact, it should give approximately the correct amount of thrust at the base of the dome; moreover, the available data do not justify a more exact computation.

SUMMARY

(1) The McCartys flow is approximately 30 miles long and several miles wide near its source; it has an area of approximately 119 square miles. The last 6 miles is between 30 and 40 feet thick. If the average thickness of the whole of the flow is taken as 75 feet, the volume is approximately 1.7 cubic miles.

(2) The flow is in the main unweathered, uneroded, bare, and barren. The Acoma Indians have a legend that the flow covers land their ancestors tilled. Pueblo I potsherds have been found in a valley fill which is probably buried by the flow. This indicates that it was extruded within the last 1200 years.

(3) Because the flow is not eroded, minor features such as grooved lava, squeeze-ups, ropes, spatter cones, tree molds, cracks, banded lava, and cavities are still present.

(4) The extrusion of the McCartys flow apparently diverted drainage from the Rio San José into the Rio Salado and Little Colorado. The Rio San José is in Atlantic drainage, the Little Colorado in Pacific.

(5) The presence of flow units, revealed physiographically, of tongues at the end of the flow, and of miniature flow units, and the orientation of the ropes, all indicate that in places the flow moved by the flow-unit mechanism. The varying thickness of the flow indicates that its velocity varied from place to place.

(6) The distribution of tree molds indicates that the climate at the time of extrusion was similar to that of the present.

(7) In an area approximately 2 miles long, near the terminus of the flow, there are about 100 collapse depressions. The largest depression is nearly a mile long and in places more than 300 feet wide, whereas the smallest is only a few feet in diameter. These depressions have been formed by the collapse of the roofs of lava tunnels. Analysis of the probable strength of the roofs of lava tunnels indicates that they owe their origin not to the collapse of one large tunnel but to the partial collapse of several small tunnels. Although many of the depressions may have been formed quite recently, in some cases collapse occurred while liquid lava was still present and the under side of the crust was still soft and plastic.

(8) On the last mile of the flow pressure ridges are very common. The shortest are about 130 feet long while the longest are more than 1200 feet. They are from 10 to 25 feet high and as much as 100 feet wide. In transverse section the sides of the pressure ridges are steep, resembling the gable of a house or the cross section of a broken anticline. They have a medial crack, running along the crest of the ridge, which may be as much as 15 feet wide. They are, in general, parallel to the flow, and close to its margin. That these pressure ridges were formed when liquid lava was still present below the crust is proved by the fact that some of the medial cracks

are filled in part with secondary lava which welled up into them after the formation of the ridges. It is suggested that the collapse of a dome between 27 and 38 feet in height, due to withdrawal of lava from beneath it, resulted in the compressional forces which buckled the crust and produced the pressure ridges.

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TUFTS COLLEGE, MEDFORD, MASSACHUSETTS

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COMMITTEE WORKING ON A NEW ROCK-COLOR CHART FOR FIELD USE

A committee representing a number of geological societies and organizations has begun work on a new rock-color chart designed specifically for field use. The membership of the committee is as follows:

Parker D. Trask, representing The Geological Society of America.

Ronald K. DeFord, representing the American Association of Petroleum Geologists.

Joseph T. Singewald, Jr. and R. M. Overbeck, representing the Association of American State Geologists.

Olaf N. Rove, representing the Society of Economic Geologists.

E. N. Goddard, representing the U. S. Geological Survey.

The first meeting of the committee was held on May 2, 1946, in Washington, D. C. Hugh D. Miser, of the Geological Survey, who had been instrumental in getting the work started, gave a brief account of the discussions and correspondence that led up to the organization of the Committee. Mr. DeFord, who was unable to attend, sent a letter suggesting a general plan of procedure, and this letter was used as a basis for discussion. The following plans were agreed upon by the committee.

1. The rock-color chart is to be based on the Munsell color system, the most widely accepted system of color identification in the United States.

2. Simple color names of the ISCC-NBS (Inter-Society Color Council-National Bureau of Standards) method are to be used on the chart, insofar as is applicable to field use. This method has already been adopted by a large number of societies and organizations interested in color.

3. In addition to the color names, the Munsell hue, value, and chroma designations are to be put on the chart, for the use of any geologists who feel the need of numerical designations and fine color distinctions.

4. Sedimentary, igneous, and metamorphic rocks (both consolidated and unconsolidated) are to be included, and also well cuttings. If possible, both wet and dry rocks are to be included.

The committee is now engaged in collecting and classifying the widest possible range of rock specimens in order to determine the range of colors needed on the chart. The next meeting of the committee is to be held at the Chicago meetings of The Geological Society of America in December 1946, and at that time the following problems will be considered:

1. The number and range of colors and the appropriate names to be used on the chart.

2. The shape, size, and arrangement of the chart, the size of the color tabs, and whether holes in the chart should be used to facilitate comparisons with rock samples.

3. The method and means of printing and distributing the chart.

The committee will welcome any suggestions. The Munsell color system, as applied to rock colors, is discussed by DeFord (*Rock colors*, Am. Assoc. Petrol. Geol., Bull., vol. 28, no. 1, p. 128–137, January 1944). The ISCC-NBS method of naming colors is described by Dean B. Judd and Kenneth L. Kelly (*Method of designating colors*, Nato. Bur. Standards, Research Paper no. 1239, September 1939).

OFFICIAL CALENDAR

December 26–28 Annual Meeting of The Geological Society (with the Mineralogical Society of America, the Paleontological Society, Society of Vertebrate Peleontology) at The Stevens, Chicago, Illinois.

December 30–31 Meeting of Section E, Am. Assoc. Adv. Sci., at the Institute of Geographical Exploration, Harvard University.